V. BELOUSOV

structural geology

MIR PUBLISHERS

В. В. БЕЛОУСОВ

СТРУКТУРНАЯ ГЕОЛОГИЯ

ИЗДАТЕЛЬСТВО
МОСКОВСКОГО УНИВЕРСИТЕТА

Structural Geology

BY

V. V. BELOUSOV

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A. GUREVICH

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Contents

Introduction

Structural geology is the division of geotectonics concerned with the modes of occurrence of rocks in the earth's crust. It is known from studies of the structure of the earth's crust that rocks form bodies of different shapes. For example, sedimentary rocks generally occur as beds or strata, that is, bodies of great areal extent but of limited thickness. The beds may be horizontal, tilted or bent into convex and concave folds. Igneous intrusive bodies occur in the earth's crust as dome-shaped, cylindrical, drop-shaped or in other forms.

The bodies formed in the earth's crust by various rocks are referred to as structural forms. Sometimes they are simply called "structures" but this usage is not exactly correct. Therefore, one may speak for instance of a "folded structure" in reference to the structure of a particular area of the earth's crust characterized by the presence of folds, but it would be wrong to apply the term "structure" to a single fold, as it is often done. In this case the proper term would be "structural form". Study of the modes of occurrence of rocks or structural forms is the subject of structural geology.

The modes of occurrence of rocks may be studied morphologically, when the objects of the study are the forms in which mineral bodies occur, or they may be studied genetically when the object is the origin of a particular structural form. The principal object of structural geology is the morphological study of the modes of occurrence of rocks, that is the description of the outward appearance of various structural forms and their classification. However, a purely morphological classification of structural forms, that takes no account of the conditions under which they had developed, cannot be satisfactory, since it may result in the lumping together of structural forms of widely different origin which will make it practically worthless. Therefore, structural geology also concerns itself with the origin of structural forms.

Furthermore, the conditions that give rise to various structural forms may be studied from different points of view. For example, the problem may be a kinematic one, the object being to ascertain the nature of the movement of the earth's crust that directly resulted in the development of a given structural form. This is often referred

to as the study of the "mechanism" of development of structural forms. Thus, investigating the folds of a certain stratum, the researcher establishes how the material of the stratum had moved in the process of development of the given fold.

Structural forms may also be studied in the historical aspect. The object would be to establish the succession of geologically different structures through geological time.

Studies of structural forms may also pose problems of a genetic or dynamic nature: what forces gave rise to the given structural form, in what direction they acted, and what was their origin. Under various conditions the same forces may give rise to different structural forms. Therefore, when the genetic aspect of the structural forms is considered, it is necessary to study the relationship between the forces which acted upon the rocks and the character of rocks, the original bedding of the strata, their depth beneath the earth's surface, etc.

Structural geology concerns itself with kinematic problems incident to the study of morphological features of structural forms. As far as historical and genetic problems are concerned, they belong principally to the domain of general geotectonics.

The importance of structural geology as a part of the general system of geological sciences can hardly be overestimated. Good geologic surveying is impossible without proper understanding of structural forms. Unless the geologist understands what may be the mode of occurrence of rocks in a given case, unless he is able to correlate various rock exposures, his map will be structurally mute. It would consist of a hotch-potch of irregular spots from which no practical conclusions can be drawn.

Geologic mapping consists of revealing, studying and recording on maps and profiles the structural forms in the area in question. Therefore, at Soviet higher schools structural geology is taught in close relationship to the course of geologic mapping.

The modes of occurrences of rocks have a fundamental effect on the distribution of all economic minerals. For example, oil and natural gas as a rule concentrate in the crests of anticlinal folds; therefore, to find oil one must first locate anticlinal folds. This is hard to achieve without good understanding of all the morphological features of folded structures. In view of that the oil geologist uses particularly delicate methods for investigating the modes of occurrence of rocks. Many ore minerals of deep-seated origin form ore bodies that fill fissures in the earth's crust. Structural geology provides data needed for understanding the specific features of fissures of different kinds and the dependences controlling their distribution in the earth's crust. Applied geology has a special division called "the structure of ore deposits".

Knowledge of the modes of occurrence of rocks is very important for solving hydrogeological problems, inasmuch as the pathways and the manner of movement of subterranean waters are controlled by them. The modes of occurrence of rocks must be taken into account in geological surveys for civil engineering projects.

Studies of the conditions of occurrence of rocks provide the foundation for theoretical conclusions on the history of development of a given segment of the earth's crust and on the tectonic movements that had occurred here. Knowledge of structural forms is as important to the geologist as knowledge of anatomy to the physiologist. Possibly knowledge of morphology is even more essential to the geologist considering that the physiologist can observe certain particular functions of the human body directly, whereas the geologist can only reconstruct the history of tectonic movements of the earth's crust from their ultimate results as recorded in the mode of occurrence of rocks.

In describing various structural forms in the subsequent discussion we shall not dwell on the methods of study thereof, inasmuch as the principal methods of studying the modes of occurrence of rocks are treated in the course of geologic mapping. We shall briefly describe only certain special methods of structural geology. Even then we shall not go into the details, but confine ourselves to stating their principles.

The course of structural geology relies on the information received by the student in the course of physical geology and also during field practice when he gets a general idea of the occurrence of rocks in nature.

CHAPTER I

Primary Structural Forms

General

The structural forms of rocks may be divided into primary and secondary. Primary structural forms are those that develop in the process of the formation of rocks and are closely connected with the conditions under which they had formed. Secondary structural forms develop as a result of various later changes in the primary forms. Such changes are generally caused by mechanical factors of different nature, among which the tectonic movements of the earth's crust play the dominant role. Therefore, in most instances secondary structural forms result from changes in the original mode of occurrence of rocks under the effect of tectonic movements. Secondary tectonic structural forms are also referred to as tectonic disturbances or tectonic dislocations.

As a rule, the primary structural forms of sedimentary and igneous rocks are different. Metamorphic rocks which are the product of alteration of either sedimentary or igneous rocks in some cases show structural forms characteristic of sedimentary rocks and in other cases those typical of igneous rocks. Considering, however, that metamorphism generally develops in areas of intensive tectonic movements, metamorphic rocks as a rule show only secondary structural forms.

Apart from the structural forms characterizing a particular rock, the internal arrangement of its constituents (grains, crystals, fossils) must be also considered. Different terms are used to describe the specific features of the internal structure of rocks. For example, the term *structure* is often applied to features associated with the form and size of grains, whereas the features related to the mutual arrangement of the grains are referred to as *texture*.

The primary internal structure of a rock is controlled by the conditions of its formation. The primary structure, however, may change in the course of subsequent tectonic phenomena, resulting in the development of a secondary internal structure. Changes in external structural forms caused by tectonic factors are referred to as tectonic disturbances or dislocations. Changes in the internal structure of rocks caused by tectonic factors may be referred to as internal tectonic dislocations.

1. PRIMARY STRUCTURAL FORMS OF SEDIMENTARY ROCKS

The Stratum as a Structural Form

The most common primary structural form of sedimentary rocks is a horizontal stratum or bed.

A stratum or bed is a relatively thin body of sedimentary rock of great horizontal extent. Its thickness may range from a few centimetres to several metres, while horizontally a stratum may extend for hundreds of metres, even several kilometres or more.

In stratified sedimentary series, as a rule in the vertical direction, alternation of beds of different composition is observed. For example, a stratum of coarse sandstone may be overlain by a layer of fine-grained sandstone, followed higher up by a layer of clay after which comes a layer of marl and then sandstone again, etc. However, stratification is observed also in homogeneous sedimentary rocks. For example, a homogeneous limestone series is always divided into different layers. In such cases the layers are bounded at the top and bottom by visible horizontal partings. Similar partings are observed between layers of different composition.

The stratum, considered as a structural element of sedimentary series, has a *floor* and a *roof*. The roof of the stratum is the *bedding plane* for the overlying stratum.

In the vast majority of cases the primary position of strata is horizontal. This is explained by the conditions of their formation, since strata are usually deposited either on the floor of shallow seas levelled by abrasion or on the surface of low continental plains and valley floors levelled by subaerial processes.

In some instances, however, the primary position of strata may be tilted. This is observed on valley slopes, on steep submerged slopes of seashores, at the fringes of reef coral masses or when the sediments fill cavities in the roof of underlying rocks. Most often primary inclined attitude of strata is observed in recent or very young, geologically speaking, sediments (Quaternary and Upper Tertiary). In older sediments this is observed less often because with time tilted strata that had formed on relatively steep sections of the sea floor or in valleys were easily destroyed by later marine abrasion or subaerial erosion. After that the horizontal strata deposited on top of the levelled surface.

Primary tilting of strata is observed in limited stratigraphic intervals, being rapidly replaced by horizontal bedding higher up the section (Fig. 1).

The angle of primary tilting of strata rarely exceeds 15 to 20°. Thus, in most cases strata originally are horizontal.

The roof and the foot of a stratum may be flat and parallel, though occasionally they may be undulating or show an irregular relief with projections and depressions.

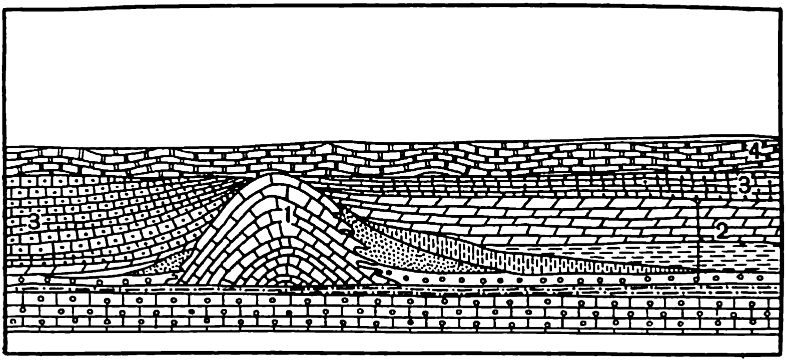

Fig. 1. Primary inclined bedding of strata. Lower Carbon reef mass in New Mexico, USA:

1—Alamagordo reef mass; *2*—stratified Alamagordo rocks; *3*—Arcente suite; *4*—Donna Anna suite (after Laudon)

Flat bedding planes are commonly observed where sedimentation was continuous. Uneven surfaces of strata most often bear evidence to an interval in the deposition of sediments, during which the earlier-formed strata suffered erosion and scouring.

Primary Internal Structure of Sedimentary Rocks

Since the internal structure of sedimentary strata is controlled mainly by the physico-geographic conditions of deposition of the given rock it displays great diversity. These problems are considered in detail in the course of petrography of sedimentary rocks and also in the course of lithology. We shall mention here briefly the main features of the internal structure of sedimentary strata.

These features primarily depend on the mutual arrangement of the grains in a given rock. This arrangement is controlled by the movement of the medium in which the sediment was deposited and is expressed best in coarse-fragmental material (conglomerates and sands). For example, in near-shore marine conglomerates one often observes an imbricate arrangement of flattened pebbles. They are inclined mostly in the direction from which came the surf waves, that is towards the sea, while the long axes of elongated pebbles most often are parallel to the shore line. In river channels pebbles usually lie at right angles to the stream and are inclined against the current.

The orientation of elongated grains of sedimentary rocks, such as long shells, is also controlled by the movement of water. In river streams they line up in the direction of the current; furthermore funnel-shaped shells are usually so oriented that the funnels point downstream; in the surf zone elongated shells align themselves

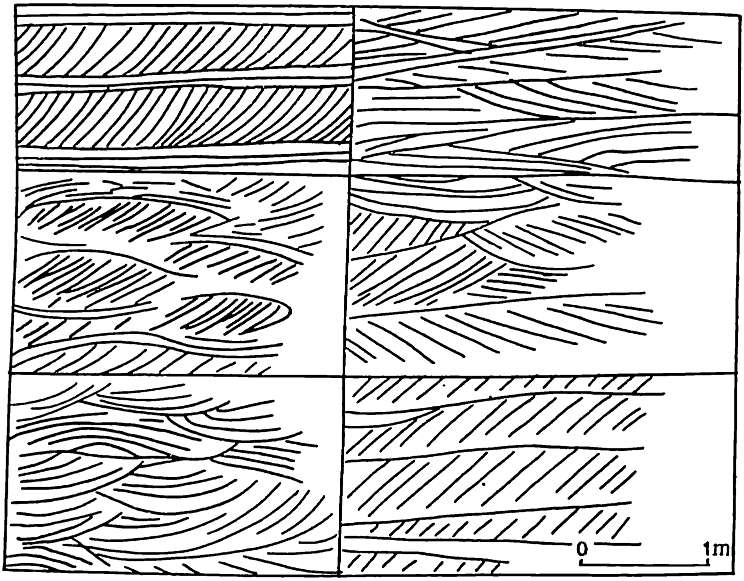

Fig. 2. Different kinds of cross-bedding

parallel to the shore while the funnels are randomly oriented. Elongated grains in sands also show a definite orientation, mostly in the direction of the stream. This also applies to marine sands deposited in the zone of strong currents.

On detailed examination of a stratum one usually observes a kind of stratification of a smaller scale represented by extremely thin laminae from one to a few grains thick. Such laminae are distinguished by different size and composition of the grains and often by colour. They are observed in conglomerates, sands, clays, less often in limestones. The laminae may be parallel to the entire stratum. Furthermore, they may be band-shaped traceable over large distances, or lenticular, discontinuous and rapidly pinch out. Often there is observed microlamination due to inclusion of mica, clay gouges and vegetable detritus in the homogeneous sandy rock. The laminae may also be tilted relatively to the stratum. This is referred to as cross-bedding. Examples are shown in Fig. 2. Generally, near the

floor of a stratum the tilted lamina flattens out assuming a near-horizontal attitude, parallel to the stratum.

Cross-bedding develops when sand is deposited on the bottom of river streams, in the surf zone and in dunes. Several varieties of cross-bedding are recognized. Their character indicates the conditions under which the sands had formed. It may be used also for distinguishing the roof and the floor of strata in strongly dislocated rocks.

Intercalations in sand beds may be wavy which is usually the result of the appearance of ripples on the surface of the sand under the action of waves in very shallow water or under the action of wind. Sometimes the intercalations in clay or argillite layers show very complicated bent and twisted forms due to the slumping of silty water-saturated sediment (during its formation) along the floor of a basin with a very gentle gradient.

A most important feature of the internal structure of a stratum is the presence of organic fossils. It is well known that the geologic age of sedimentary rocks is determined by fossils. They also help to ascertain the physico-geographic conditions under which these rocks had formed. Fossils are important to structural geology as well. If the structural geologist finds remains of burrowing organisms which should be in a definite position relatively to the floor, he can determine the position of the roof and floor of the stratum; distortion of the original form of the organism (flattening, extension, rupture) is an indication of the degree and character of the tectonic deformation suffered by the stratum.

Combinations of Strata

Formations. Only in rare instances the alternation of rocks associated with stratification seems to be haphazard. Usually a certain regularity is observed in the stratigraphic position of rocks. For example, one part of the section of considerable thickness may be composed of a clay series, in which other rocks (sandstones, limestones) form thin and rare intercalations, another part of the same section may be predominantly composed of limestones or constitute a regular alternation of limestones, clays, etc. Such rock sequence makes it possible to identify what is usually referred to as *sedimentary rock formations*. Thus, a formation is a sequence composed of alternating rocks which are in a certain regular quantitative relationship among themselves. This concept of a formation is a morphological one. Geotectonics is concerned with a more general concept of formation, based on the tectonic conditions under which they had formed. The designations given to different formations reflect the rocks of which they are predominantly composed, thus one speaks of limestone formation, slate formation, sandclay formation, etc.

A specific kind of combination of layers are so-called *rhythmic sequences*, characterized by repetition at regular intervals of definite rock sequences. Such a sequence, from the bottom upward, may

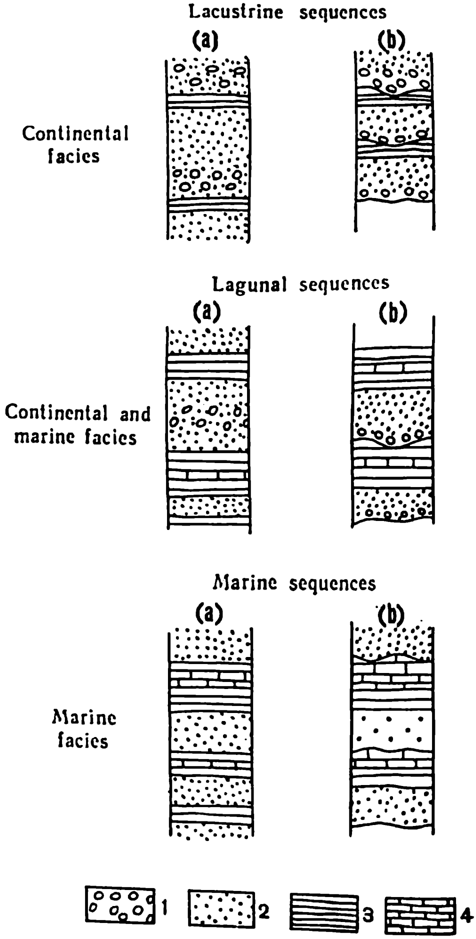

Fig. 3. Examples of regular alternations:
a—continuous; *b*—discontinuous (after L. Rukhin); *1*—conglomerates; *2*—sands; *3*—clays; *4*—limestones

consist of coarse-grained sandstones, clay marl, and limestone. Each sequence is usually not more than 2 or 3 m thick. Hundreds of thousands of such sequences follow one another in the section, differing among themselves in thickness (of the whole sequence or of its individual elements). The sequences also vary areally in thickness and composition. Regular alternations are most typical

of flysch formations usually composed of alternating bands of sandstone or limestone, marl or clay. Flysch formations are characterized by great thickness (a few kilometres). Regular alternations are also often observed in coal measures, in which continental deposits are repeated in this order from the bottom upward: sandstones, aleurolites, clays, coal and marine deposits (clays, limestones, sands). The boundaries between the sequences are always sharp, while the transition between the elements within a sequence is gradual.

These regular alternations can also be helpful for identifying the roof and the floor of a bed, which is most important when the strata are reversed. Such identification is made possible by the fact that in most cases the coarser sediments occur at the foot of a sequence, whereas the roof is characterized by finer sediments; further up there is observed a sharp transition to coarser rocks of the next sequence (Fig. 3).

Different forms of stratifications, the changes in formations from the bottom upward or from top to bottom reflect the vertical variability of sedimentary series. Changes in the lithological composition of sedimentary rocks are usually referred to as facial changes (changes of facies or lithofacies). This refers to the fact that beds of different ages following one another differ in composition or, using the generally-accepted term, in their *lithofacies.*

Areal Variations in Lithofacies and Thickness of Sediments

An important characteristic of sedimentary series is that they vary not only vertically but horizontally as well. This is to say, that as one moves from one point of a given area to another, one notices that series of a certain age change both in composition and thickness. In different places, series of the same age may be composed of different rocks characterized by varying thickness.

The horizontal variations may be differently expressed in each specific case. In some regions the changes are gradual and become noticeable only over vast stretches (hundreds of kilometres); in other instances they are sudden and distinct over short stretches. Such sharp changes are observed in geosynclines, whereas platforms are characterized by strata of sustained composition and thickness (see Chapter VIII).

Figure 4 shows an example of horizontal variation in lithofacies and thickness of a sedimentary series. In the zone of transition of lithofacies, pinching of beds of one composition and the appearance of beds of a different composition can be observed. In view of that at the boundary between two lithofacies the beds of one facies are usually interfingered with those of another. Considering that with

time the zone of change of lithofacies is often displaced in some direction, the boundary between them is rarely vertical. Most often it is inclined; furthermore the direction of the incline may vary.

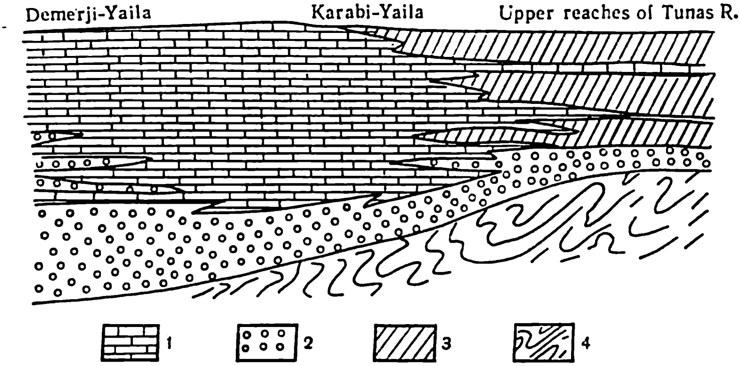

Fig. 4. Schematic representation of thickness and lithofacies variations of Upper Jurassic strata in the Crimea (after M. Muratov): *1*—limestones; *2*—conglomerates; *3*—flysch (clays, marls, limestones); *4*— "Taurian" clay shales

Occasionally a certain lithofacies has a very limited extent, occupying, for example, a narrow band between rocks of different lithofacies. If such a lithofacies had formed within a very limited space

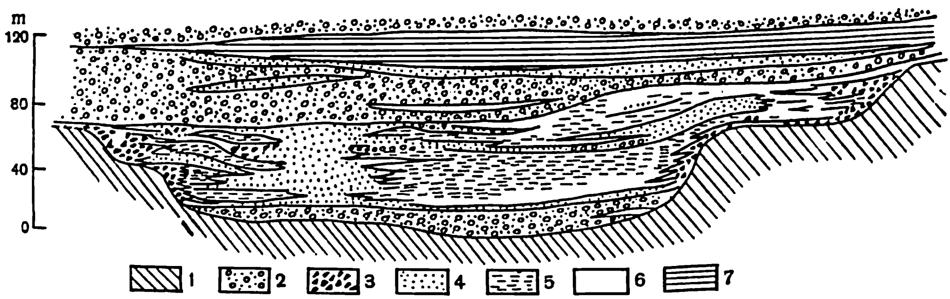

Fig. 5. Lenticular occurrence of rocks at the northern fringe of the Moscow coal basin (after E. Bruns): *1*—Paleozoic rocks; *2*—conglomerates; *3*—breccias; *4*—sandstones; *5*—sandstones and clays; *6*—clays; *7*—coals

of time, its rocks may appear in the vertical section as short lenses rapidly pinching out in two opposite, or any other, directions. Such lenticular mode of occurrence is most typical of sedimentary rocks of continental origin (river sands, for example) (Fig. 5).

Changes in the thickness of sedimentary series are usually accompanied by gradual pinching of one layer after another, or conversely by the appearance of new layers with a corresponding decrease or increase of the total thickness of the series.

Causes of Thickness and Lithofacies Variations

Variations in the thickness and facies of sedimentary series reflect the process of slow upwarping and downwarping of the earth's crust. The thickness of sedimentary series corresponds to the magnitude of the subsidence of the earth's crust. In geological sections we observe either predominantly shallow marine sediments or deltaic, or other continental near-shore deposits. Deposits of these kinds accumulate in regions of subsidence of the earth's crust and, as it were, make up for this sagging. Therefore, the greater is the sag of the earth's crust the thicker are the sediments accumulating in it. Thus if in one place the Jurassic strata are 300 m thick while in another they have a thickness of 2,000 m, that means that in the Jurassic time the earth's crust sagged in both places, but in one place the subsidence amounted to 300 m and in the second to 2,000 m.

Where the earth's crust was upwarped, its surface rose above the sea level and above the level of continental accumulation and suffered erosion and denudation. The erosion products are carried down into the adjacent depression where they are deposited in zones. The coarse fragmental material is deposited near the shore and the fine detritus farther out. Limestones and other sediments, which are associated with an influx of detrital material only in an insignificant degree, are deposited still farther off-shore. This is the mechanism responsible for the formation of different lithofacial zones and for areal variations in lithofacies of the same geologic age.

As upwarping of land becomes more intensive, the quantity of detrital material increases and it is distributed over a still greater area. The rise of new areas of land and disappearance of older ones as a result of subsidence causes a change in the distribution and character of the lithofacics with time, that is vertically in the section. For example, regular alternations most often reflect the periodic intensification and quiescence of subsidence in the adjacent area suffering erosion.

Massive Occurrence of Sedimentary Rocks

In some cases sedimentary rocks occur originally not as beds but as more or less large masses. Such a mode of occurrence is typical of reef limestones. These are formed mostly by colonial corals, but

often also by algae, bryozoa and archeocyathens. Massive accumulations of these colonial organisms are referred to as *bioherms.*

Bioherms may be dome-shaped, roughly-lenticular or mushroom-like. They are usually surrounded by layers of fragmental recrystallized limestones largely derived from the disintegration of the reefs. The layers are often inclined outward from the reef at an angle of 10 to 15° (see Fig. 1).

2. *PRIMARY STRUCTURAL FORMS OF IGNEOUS ROCKS*

Structural forms of effusive and intrusive igneous rocks vary with the conditions of their formation.

Effusive Rocks

The primary structural forms of effusive rocks bear some resemblance to those of sedimentary rocks. Lavas that possess great fluidity (basic and intermediate) spread out to form flows or sheets of varying thickness. In the case of repeated effusions when the last lava flow had time to congeal before the next sheet was formed, sequences composed of many flows of varying thickness and composition are formed.

Very often lava flows alternate with beds of volcanic tuffs and breccias and also with beds of sedimentary and mixed volcanogenic-sedimentary rocks.

Depending on the volume of lava effusions and fluidity of the lavas, the area of lava flows may differ in the range within wide limits. Basic lavas spread over vast areas and form sheets of surprisingly uniform thickness. More viscous acid lavas spread over smaller areas and often form dome-shaped accumulations over the centres of effusion, and are referred to as *extrusive domes.*

Where denudation exposes the vent funnel of an ancient volcano, effusive rocks in the form of the so-called necks or volcanic pipes are observed. They are filled with lava or breccia composed of frozen lava fragments. The necks occur as irregular cylinders, pipes or lenses and often are a few kilometres across.

The elements of the primary internal structure of effusive rocks are various blocks which develop in the process of formation of the effusive rock and also flow lines. Regularly-shaped blocks are bounded by joints that develop as a result of the cracking of lava when it solidifies and, therefore, shrinks in volume.

Most common in basic lavas is the so-called *columnar* or *prismatic jointing* by which the frozen lava flow is split into polyhedral prismatic columns which are generally perpendicular to the surface of the flow.

Also known is *globular* or *hammock* jointing believed to be caused by rapid cooling of lava on contact with water.

Finer elements of internal structure of lava may be represented by fluidal texture expressed in the orientation of individual elongated *phenocrysts** and also gas vesicles and amygdaloids in the direction of the lava flow (Fig. 6). On the basis of fluidal texture it is possible to deduce the direction of ancient lava flows.

Other features of the primary internal structure of effusive rocks are variations in composition, texture and appearance often observed between the floor and the roof of a single lava flow. When

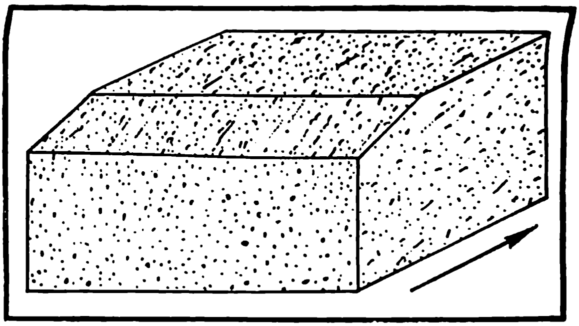

Fig. 6. Fluidal texture. The arrow indicates the direction of the movement of the lava flow

a lava flow congeals the gas bubbles collect in its upper part in view of which the upper part of the solidified rock generally contains more cavities and amygdaloids. Often near the roof the flow is more acidic than at the floor. Closely associated with lavas are layers of tuffs and breccias. In typical tuffs and breccias, their constituents—solid volcanic products—are randomly distributed. If the pyroclastic material was to some extent transported by water or redeposited by surf, it is mixed with some purely sedimentary material and such rocks of a mixed volcanogenic-sedimentary origin (tuffites) greatly resemble sedimentary rocks and, in particular, show definite stratification.

Effusive rocks are grouped in the earth's crust into more or less large complexes called formations, similar to sedimentary rock complexes. All volcanogenic formations are characterized by the predominance of lavas or pyroclastic material, a definite chemical composition of the rocks, and a definite alternation of lavas, tuffs and breccias.

* Phenocrysts (or phenocrystals) are analogues of porphyry segregations, i.e., more or less large crystals or crystalline grains occurring in fine-grained semi-vitreous or vitreous groundmass of porphyry rocks.

Occasionally the occurrence of volcanogenic rocks displays a certain regular pattern, such as alternation of lavas of different composition (basic, intermediate and acid), alternation of lavas and tuffs, or of tuffs and breccias of different composition and coarseness. Sedimentary rocks regularly alternating with volcanogenic may also form part of this regular pattern.

Naturally, the character and composition of effusive rocks may vary not only vertically but horizontally as well. As in the case

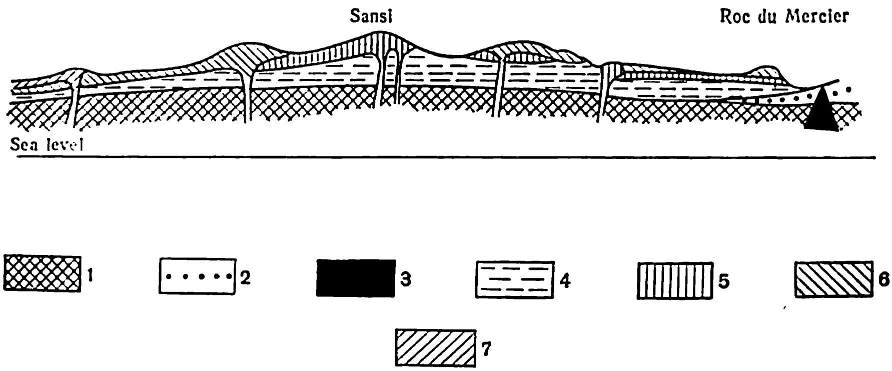

Fig. 7. Variation of volcanogenic rock facies. The Mont D'Or massif (France): 1—basement; 2—rhyolite tuff; 3—phonolite (effusive analogue of nepheline syenites); 4—andesite tuff; 5—porphyritic trachyte; 6—hornfelsitic andesite; 7—basalt

of sedimentary rocks, the composition of effusive rocks, that is their "facies", is subject to areal variations which means that within a single stratigraphic level lavas may be replaced by tuffs, breccias, etc. (Fig. 7).

Intrusive Rocks

Intrusive rocks exhibit a much greater diversity of structural forms. The attitude of intrusive rocks relatively to sedimentary or metamorphic host rocks may be concordant or discordant (cross-cutting). In the first case the intrusive body is flat or lenticular and occurs between sedimentary strata concordantly with them. In the second, the intrusive body cuts across the enclosing strata.

This division is a conventional one. Some intrusions are partly concordant (usually at the roof) and partly cross-cutting. Therefore, one may speak of partly concordant or partly discordant intrusions.

Conformable intrusions are sheet intrusions or sills, laccoliths, composite laccoliths, lopoliths and phacoliths.

Partly concordant intrusions are intrusive domes and magmatic diapirs.

Discordant or cross-cutting intrusions are fissure intrusions, batholiths and stocks.

It should be noted that a close relationship exists between the chemical composition of intrusions and their structural forms: a definite rock composition (basic, acid, alkaline or intermediate) is characteristic of intrusive bodies of a definite shape.

Concordant Intrusions

Sheet intrusions are flat igneous bodies that had thrust along the bedding planes of sedimentary rocks. Naturally, a sheet intrusion must have some feeding channel, usually a deep fissure. Often,

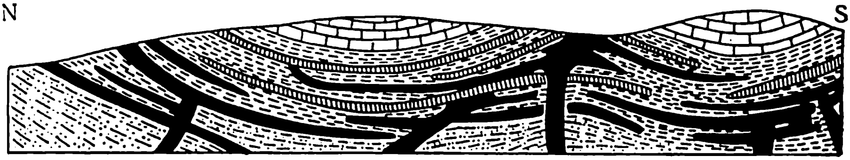

Fig. 8. Diabase sheet intrusions (black) in Ordovician and Silurian strata near Prague

however, such channels cannot be located within the area in question. Some sheet intrusions comprise thousands of square kilometres and show amazingly uniform thickness. A 300-m thick Triassic intrusion in the state of New Jersey (USA) has been traced for 160 km.

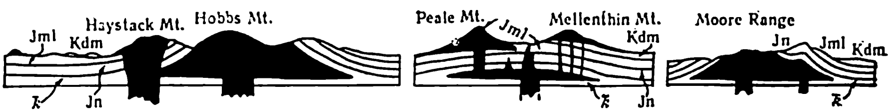

Fig. 9. Laccoliths of La Sal Mts:
Jml—Mc Elmo formation; Kdm—Dakota sandstones and Mancos clay shales; Jn—Wingate, Cayenta and Navajo sandstones; $\overline{R}$—Triassic and Permian strata

Sometimes in the same geological section a number of intrusions are interbedded with sedimentary rocks (Fig. 8). The magma of sheet intrusions is usually basic.

A special kind of sheet intrusions are *harpoliths*—bodies that have been formed along the surface of unconformity in sedimentary series.

Laccoliths are lenticular or bun-shaped bodies formed as a result of intrusion of magma between the strata and arching of the overlying strata caused by it (Fig. 9). In some cases feeding channels filled with the same igneous rock are observed as well. In other instances such channels may be missing. In the latter case the lacco-

liths should be regarded as sheet intrusions that have been pinched and broken up into separate lenses by later deformations.

Also known are complex laccoliths, composed of intrusive lenses occurring one above another in a "cedar tree" shape (Fig. 10). In

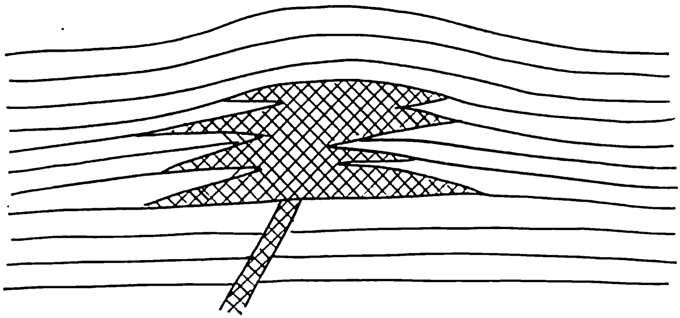

Fig. 10. "Cedar-tree" laccolith

this case the body of the laccolith as a whole cuts across the host rocks, while separate elements of this complex body are concordant with the enclosing rocks.

Laccoliths are typically composed of intermediate and alkaline rocks (andesites, nepheline syenites, etc.).

Lopoliths are huge sheet intrusions comprising many thousands of square kilometres, and are centrally depressed so that their upper

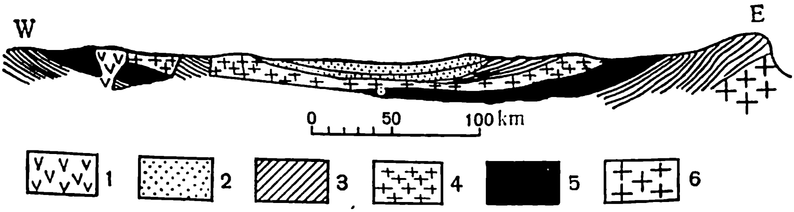

Fig. 11. Noritic lopolith of Bushveld:
1—syenites (the youngest intrusion); *2*—continental deposits of the Karoo system; *3*—Transvaal system; *4*—granite; *5*—norite; *6*—ancient granites

surface is basin-like (Fig. 11). Their composition is mostly basic with local variations up to acid.

Phacoliths are small upward or downward bent lenticular intrusive bodies located in the crests of anticlinal or troughs of synclinal folds (Fig. 12).

Phacoliths may be regarded as primary bodies only if the magma had intruded strata that were crumpled into folds earlier. Otherwise these structural forms may be the result of secondary deforma-

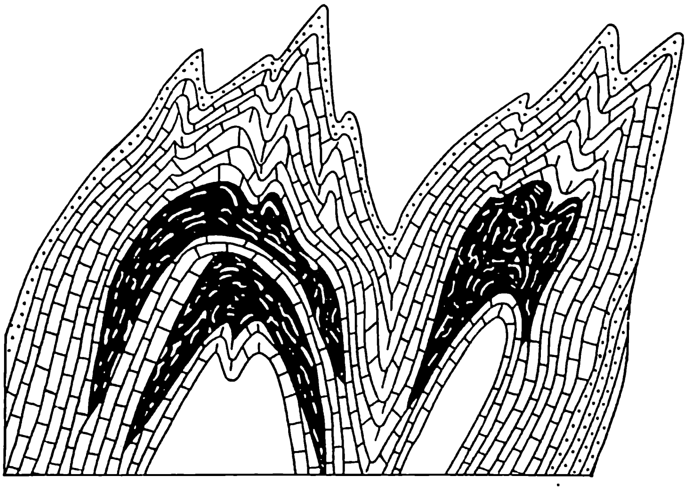

Fig. 12. Phacoliths. Black—intrusive rocks

tion of ordinary sheet intrusions that have been bent into folds together with the country rock and in this process were extended and broken up into separate large lenses.

In most instances the composition of phacoliths is basic.

Partly Concordant Intrusions

Intrusive domes and magmatic diapirs differ among themselves chiefly in size and composition. An intrusive dome may be several, even tens of kilometres in diameter. Only its dome-shaped upper

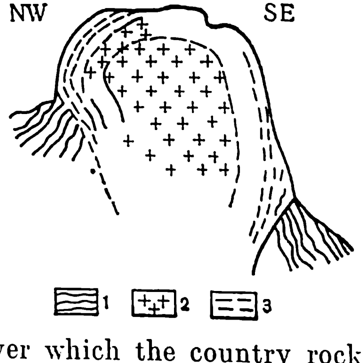

Fig. 13. Drop-like intrusion (intrusive diapir) (after V. Pavlinov):
1—country rocks; *2*—massive part of the intrusion; *3*—oriented texture at the margins of the intrusion

part usually appears on the surface over which the country rocks are also conformably arched and bent. In its deeper parts an intrusive dome is probably discordant. Intrusive domes are composed of granites and gneisses.

A magmatic diapir is a small intrusion shaped as an inverted drop, concordant with the enclosing rocks at the roof and discordant at depth (Fig. 13). Such intrusions are usually composed of intermediate or alkaline rocks.

A characteristic feature of intrusive domes and magmatic diapirs is that flat and elongated crystals within them, particularly at the edges, are oriented parallel to the outlines of the body.

Discordant Intrusions

Fissure intrusions, also called dykes and igneous veins, have formed by injection of magma into cavities that developed on the cracking of the earth's crust. Fissure intrusions are tabular bodies

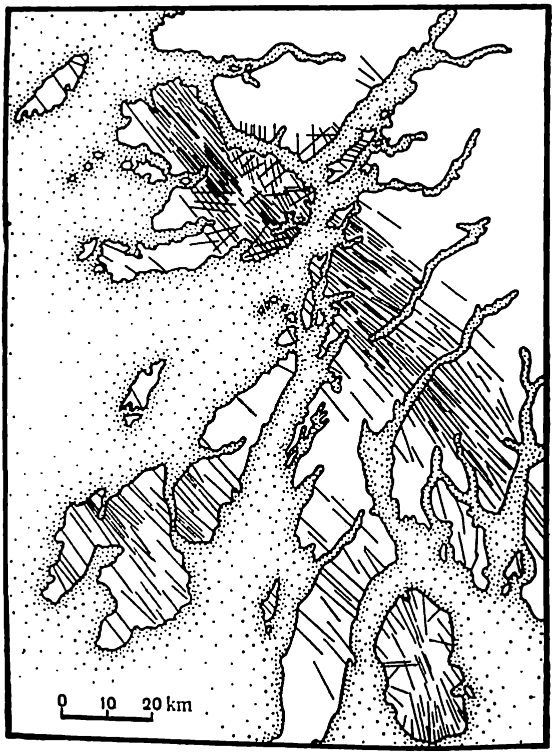

Fig. 14. Tertiary fissure intrusions of Scotland

which may be vertical or inclined. Their thickness varies from one centimetre to tens, rarely hundreds, of metres. Along the strike fissure intrusions extend for tens, sometimes for hundreds of kilometres (Fig. 14).

Most often fissure intrusions are straight, but they may also be bent and broken. A peculiar kind are systems of fissure intrusions of the cone-in-cone type in which the cones may point upwards or downwards. Such dykes usually appear on maps as circular bodies, and are referred to as ring dykes (Fig. 15). Rocks of most diverse composition from ultra acid to ultrabasic occur in fissure intrusions.

Batholiths are very large igneous bodies, the upper parts of which are dome-shaped or conical. Their form has not been fully ascertained yet since only their upper part is exposed to examination. But recently it was generally believed that these bodies were "bottomless", that they gradually widen downwards, pierce the entire crust

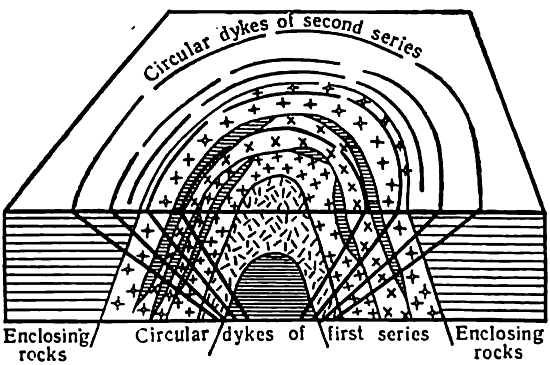

Fig. 15. Ring dykes

of the earth and directly communicate the subcrustal regions which are the source of their magma. At present many scientists believe that batholiths are specific to the upper parts of the earth's crust and at depth merge into a common granitic layer.

In the most ancient (Archean) rocks there are often observed so-called "scattered intrusions". Usually they are confined to the marginal parts of batholiths, but sometimes occupy great independent areas. Intrusions of this type give rise to migmatites—mixed rocks in which angular fragments (sometimes tens of centimetres across) of strongly metamorphosed country rocks and the enclosing magmatic matter are intimately mixed with granitic, quartzitic or aplitic igneous materials which criss-cross the entire rock in the form of countless interwining veinlets (for a more detailed description see Chapter VII).

In the horizontal section near the surface a batholith usually appears as an elongated oval the major axis of which may be hun-

dreds, sometimes thousands, of kilometres long. The chemical composition of batholiths is usually very uniform. They are composed exclusively of acid rocks, mostly granites (partly of granodiorites). As different from intrusive domes, the contact of batholiths usually cuts across the country rock (Fig. 16).

A very curious circumstance is that despite the great dimensions of batholiths they usually do not produce any mechanical effects on the country rocks. It would seem that these great igneous bodies as they intruded into the earth's crust should have pushed apart and crumpled the overlying rocks. This is not the case however.

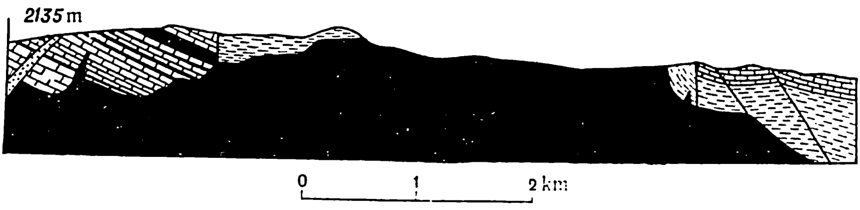

Fig. 16. The Marysville batholith (USA)

Hence the long-debated "problem of space for batholiths". The present view is that batholiths have formed not so much as a result of magmatic intrusion but through "in situ" formation of granites by recrystallization of the sedimentary shell or its "granitization" under the action of high temperature and pressure and introduction of certain highly active chemical agents (water, silica, and alkalis) from depth.

Stocks are relatively small (up to a few kilometres in diameter) irregularly shaped, often roughly cylindrical intrusions.

As different from batholiths, the country rocks at contact with stocks are strongly dislocated. Also in contrast to batholiths, the composition of stocks shows greater diversity; many of them are composed of intermediate, basic or alkaline rocks.

Intrusive bodies, especially big ones, often have irregular long and branching projections called *apophyses*.

The classification of intrusive bodies adopted in the general course of geotectonics usually differs from that used here. For example, the following groups may be recognized: batholiths (including intrusive domes and scattered intrusions), sheet intrusions (including lopoliths and phacoliths), fissure intrusions, and lastly "small" intrusions (laccoliths and stocks). Such division is most convenient since it groups together intrusions by the common tectonic conditions of their formation.

Primary Internal Structure of Intrusions

This is an important subject since the internal structure of intrusions is indicative of the conditions under which they had formed.

The internal structure of intrusions is characterized by irregular composition. Some intrusions, batholiths, for example, are very uniform in composition. But even in batholiths the marginal band usually differs in composition from the rest of the intrusion. This, so-called endocontact zone, is generally either more basic (the most often case) or more acid than the entire granite batholith); furthermore, the crystals in the endocontact zone are smaller.

In basic sheet and small intrusions this irregularity is often more pronounced. Occasionally such intrusions are distinctly divided into a basic, usually lower or inner part, and a more acid, outer part. Such a division is due to differentiation in the cooling magmatic material.

Sharp differences in composition between the central and marginal parts are also observed in fissure intrusions, which is explained by the rapid cooling of the magma near the walls of the fissure.

A curious and as yet not fully understood phenomenon of stratification is observed in basic and alkaline sheet and small intrusions. This is recorded in the fact that the material of the intrusion separates into layers from a few millimetres to hundreds of metres thick. These layers differ in composition, which may be more basic or more acid, and in the quantity of coloured mineral constituents. Such banded structure is often most pronounced in the marginal zones of intrusions, but in some instances involves the intrusion as a whole. The irregularity of composition may be explained by the fact that the intrusion in question is a multiple one, i.e., consists of several intrusions that successively forced their way into the same space.

A very important feature of internal structure of intrusive bodies is the so-called oriented texture, expressed in a regular orientation of elongated and flat crystals in the intrusive rock.

The orientation of the texture may be linear- or plane-parallel. In the first case the long axes of crystals are arranged in parallel in a certain direction, which, however, may not coincide in different parts of the intrusion (Fig. 17*a*). If the elongated minerals are also flat, their wider planes may be inclined relatively to the axis of linear orientation in any way. In the second case, flat minerals are so arranged that their wider planes are parallel to each other (Fig. 17*b*). Often these two types of oriented textures are combined; in that case both the long axes of crystals and their wider planes are parallel, which results in a linear plane-parallel texture (Fig. 17*c*).

An oriented texture develops if solid crystals have already begun to form when the magma was still moving. These crystals are deformed by the moving magma and oriented in the direction of the flow.

Closer examination of the relationship between the orientation of the crystals and the movement of the magma shows that the regular arrangement of the crystals is due not to the general movement of the magma, but to the nonuniform movement of its separate parts. The crystals may be turned about only if the magmatic flow separates into streams moving at different velocities. Then an elongated crystal caught by streams moving at different velocities will

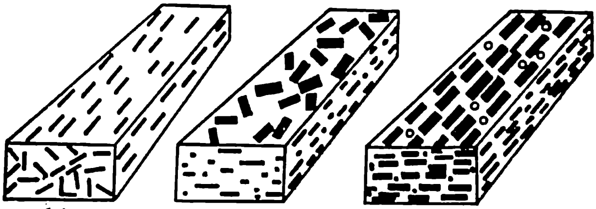

Fig. 17. Oriented textures:
a—linear; b—plane-parallel; c—linear plane-parallel

inevitably turn parallel to the stream. Nonuniform flow of the magma may be caused, among other things, by the friction of magma against the walls of the cavity it is filling. Because of that oriented textures are always more pronounced near the wall rocks, whereas the inner parts that were able to move as a homogeneous mass often do not show an oriented texture. Since the textures described above reflect the flow of magma they are also referred to as fluidal.

In crystallization occurred when the flow of liquid magma ceased, the crystals are arranged haphazardly and oriented textures do not develop.

Depending on the shape of the intrusions and the direction of the movement of the magma within them, the arrangement of the oriented textures relatively to the outlines of the intrusions may be different. Very widespread are platy flow textures bent into dome-like forms (see Fig. 13). Such dome-like internal structure concordant to the shape of the entire intrusion is characteristic of laccoliths, magmatic diapirs and intrusive domes. Occasionally, several domes of oriented textures are observed within one intrusion, which is characteristic of multiple intrusions.

In fissure intrusions, the oriented textures are either parallel to the walls or arched (Fig. 18).

Often, xenoliths, that are fragments of the country rocks, form part of the internal structure of intrusions. If the intrusive body had formed *in situ* by remelting or re-crystallization of sedimentary or metamorphic rocks, the rocks composing the xenoliths may retain their original attitude which is then closely connected with that

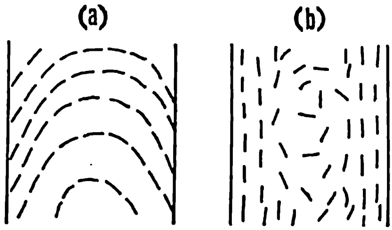

Fig. 18.
a and *b*—oriented textures in fissure intrusions (dykes)

of enclosing rocks. This results in the development of common structural forms (Fig. 19). If, however, the magma kept flowing until it solidified, the xenoliths moved together with it. They could turn around and assume a position corresponding to an oriented texture, that is with their long axes in the direction of the linear texture.

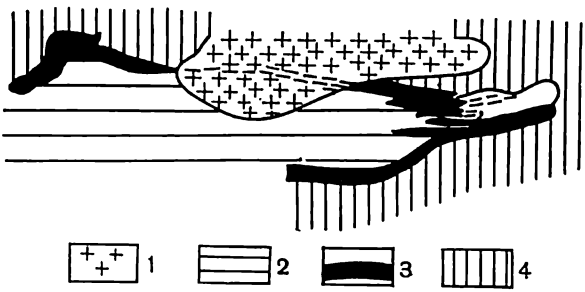

Fig. 19. Granite massifs in the Corlay coal basin (after Barrois):
1—granite; *2*—coal measures; *3*—Devon; *4*—Algonkian strata. Devonian xenoliths occur along the dotted line

The movement of the intrusive material may also be recorded in the form of *schlieren*, a term used to describe accumulations of certain minerals (dark or light) within the intrusion, as bands, ribbons or thin lenses. Such schlieren are oriented parallel to linear texture and may be composed of mica, quartz, feldspars and other minerals.

Other elements of the primary internal structure of intrusions are cracks or joints that develop in intrusive rocks upon cooling and

contraction. Such joints usually split rocks into regular parallelepiped blocks, though irregular, for example, hammock, oval, globular or other kinds of jointings are known as well.

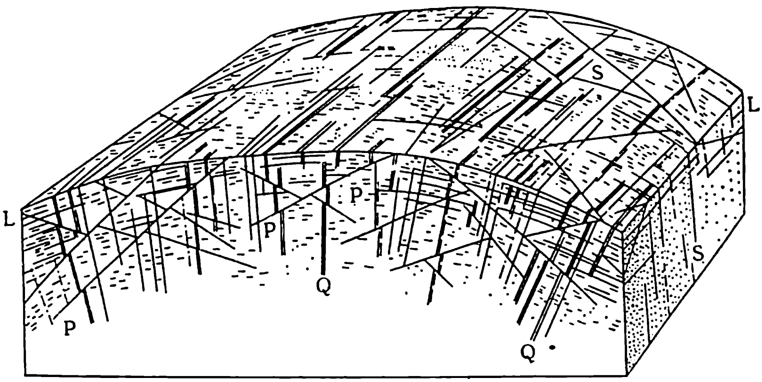

Fig. 20. Systems of joints in a granite mass (after Cloos). Oriented texture is shown by dashes

The joints bounding the parallelepiped blocks are usually classified by their attitude relatively to the oriented textures. If they run across the linearly oriented textures they are called transverse

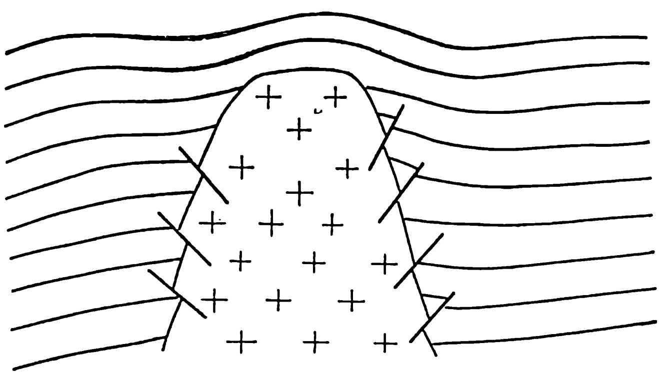

Fig. 21. Marginal faults in an intrusion

joints (Q-joints). Joints that on plan are parallel to the linear texture and usually dip at a high angle are termed longitudinal (S-joints). Low-dipping joints parallel to the linear texture (L-joints) are sometimes referred to as sheet joints. There are also recognized diagonal or oblique joints (P-joints) that bisect the angle between Q- and

S-joints (Fig. 20). This classification has been suggested by the German scholar Cloos.

All kinds of jointing described above are closely related to oriented textures and most often occur where such textures are strongly pronounced, for instance in the marginal zones of intrusive masses. The conditions under which these joints are formed will be examined in detail in Chapter VI.

The so-called marginal faults or overthrusts are often also classed among the primary elements of the structure of intrusive masses. Presumably they develop within a mass that has already solidified at the edges, but continues to move under the pressure of the still fluid magma of its inner parts. The friction developing between the upward moving frozed shell of the mass and the country rock results in shear deformation (see page 36). Shear results in the development of tension and shear fractures and also in the displacement of rocks along these fractures. Both the marginal part of the intrusion and the country rock may be involved in these movements (Fig. 21).

3. METHODS OF INVESTIGATION OF PRIMARY STRUCTURAL FORMS OF IGNEOUS ROCKS

Certain special methods may be applied to the study of igneous bodies, apart from the conventional methods of geologic mapping and petrological investigation. Here we can mention briefly only some of them.

Oriented textures, which interest us only inasmuch as they help to reconstruct the mechanism of intrusion, may be roughly determined by means of a Brunton compass. The attitude of both linear and plane-parallel textures is recorded on the map by appropriate symbols. Considering that individual crystals may deviate from the average position characteristic of a given area, the values recorded on the map will represent the average of a great number of measurements. It should be borne in mind, however, that the attitude of oriented textures varies from one part of an intrusion to another, in view of which such averaging makes sense only for a limited volume of rock, confined to a definite structural setting.

A better, so-called petrostructural method of studying oriented textures consists of examining oriented polished sections under the polarization microscope by means of the Fyodorov universal stage. This method is described in detail in petrographic manuals. Essentially it consists of establishing the spatial position of optical axes (of cleavage poles in some cases) of a large number of crystals of a certain mineral (for example, quartz or mica) after which the results are statistically processed to ascertain their average position. Inasmuch as in acicular, platy and prismatic crystals, the position of

their optical axis is closely related to crystal form, this very fact determines the spatial position of the oriented textures.

The results of studies recorded on a map and profiles will give a picture of the internal structure of the intrusion and permit to ascertain the character and the sequence in which the intruding material had moved.

Primary joints in intrusions are investigated by measuring the attitude of a large number of joints, and identifying different systems of joints by statistical methods. A joint may be assigned to a particular type only in relation to oriented texture. Thus, the study of the latter should always go hand in hand with the study of jointing.

CHAPTER II

Secondary Structural Forms

General

4. CAUSES OF DISTURBANCE OF THE PRIMARY STRUCTURAL FORMS

It has been mentioned previously that the primary structural forms may be altered, deformed or dislocated by later processes of various nature. Generally they may be divided into two groups: those due to the action of deep-seated forces and those of a superficial origin. In turn, the deep-seated processes that disturb the structural forms of rocks are subdivided into tectonic and magmatic. In the first case the attitude of rocks is affected by movements of the earth's crust caused by tectonic factors, i.e., processes occurring at a great depth beneath the crust, the nature of which is still poorly understood. In the second case the attitude of rocks is disturbed by intruding magma.

Superficial dislocations are mostly due to the action of gravity operating under different conditions. The immediate cause of such dislocations may be landslides, avalanches, caving of the roof of karst and other cavities, etc.

Among all these processes that disturb the position of rocks, tectonic processes are the most important, and it is on them that we shall centre our discussion.

Superficial dislocations and the structural forms produced by them in many ways resemble those produced by tectonic processes. Superficial dislocations result in the development of folds and faults which often differ from those produced by tectonic factors in scale only. The mechanisms of the two kinds of dislocations also have much in common. Furthermore, in a broader sense, of late the dividing line between tectonic and superficial dislocations has been largely obliterated. According to modern concepts ever increasing importance is assigned to gravity as a factor causing tectonic dislocations. For example, the formation of nappes or overthrust sheets which differ from ordinary landslides only in scale is explained by downslope sliding under the action of gravity. In view of that it is sometimes difficult to draw the line between tectonic and non-tectonic dislocations.

Apparently, it would be more proper not to divide them at all but to consider them as different manifestations of a complicated process of transformation of the structure of the earth's crust, which in different cases extends to different depths. In this book we classify as tectonic those processes due to gravity that give rise to major structural forms of a "tectonic scale" and change the position of rocks to great depth, commensurable with the depths reached by human technologies. Processes that result in minor local dislocations unrelated to the general picture of the tectonic structure of a given region, which are purely superficial and develop chiefly in the zone of weathering of rocks are classified as non-tectonic dislocations and are examined separately in the last chapter.

5. THE PRINCIPAL KINDS OF DISLOCATIONS

In the case of dislocations of primary structural forms rocks may or may not lose their cohesion. In the first case the rock changes its form without loss of cohesion, that is without rupturing. In the second case, dislocation involves rupturing, that is the development of cracks or fissures.

In this connection there are recognized dislocations without a break in continuity or plastic, and dislocations with a break in continuity or faults. Dislocations of the first kind are often referred to as folding inasmuch as they mainly occur in the form of folds. The term, however, does not cover all kinds of plastic dislocations; some of them, for example boudinage (of which later) cannot be described as folded dislocations.

We shall first consider separately the principal kinds of folding and faulting. It should be noted, however, that such division is largely conventional and mainly helps to make morphological descriptions of different dislocations more systematic. Further, we shall examine certain common combinations of folds and faults.

However, before proceeding to a description of various dislocations we shall have to consider briefly certain fundamental principles of the theory of deformation and destruction of solid bodies.

6. THE MECHANISM OF DEFORMATION

A fold is the result of a plastic deformation of a rock. A fault is a manifestation of the process of its rupturing.

The theory of deformation and rupturing of solid bodies is the subject of mechanics, physics and certain applied sciences. Correct understanding of the mechanism of dislocations requires knowledge of certain elementary conclusions from these theories.

A solid body under a load, i.e., subjected to the action of external mechanical forces which are so balanced that no linear or rotary

motion is imparted to it, suffers deformation consisting of a change of either form or volume, or both.

The solid body offers resistance to the deforming forces and in the process of deformation this resistance must be overcome. This means that in the process of deformation within the body there develop forces that oppose the external forces. These internal forces related to a unit area of a certain cross-section of the body are called stresses operating on the given elementary unit area.

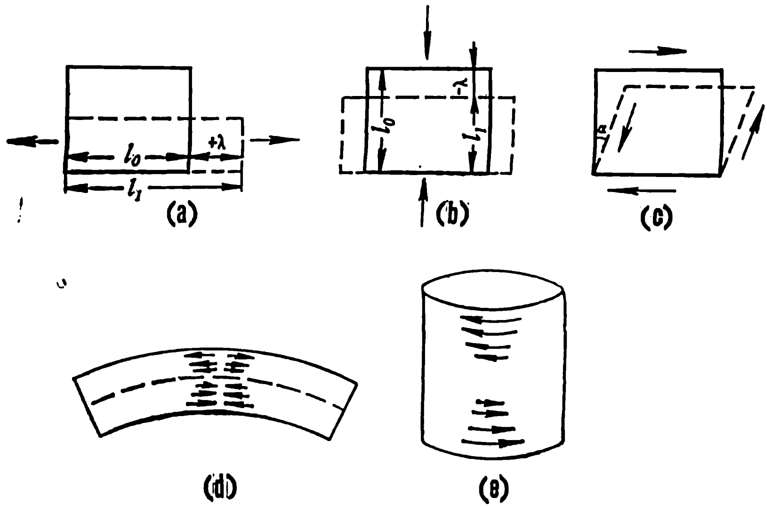

Fig. 22. Simple kinds of stresses and deformations of bodies: *a*—tension; *b*—compression; *c*—shear; *d*—bending; *e*—torsion

Depending on the direction of the applied forces, the physical body is said to be under different kinds of stresses, among which, however, several most simple kinds may be recognized: compression-extension, shear, bending and torsion. Under the action of these loads the body suffers corresponding general deformations (Fig. 22).

If, however, we turn to elementary, infinitesimal volumes of the body, we shall find, as postulated by the theory of elasticity, that any complex combination of deforming stresses applied to a body in different directions may be reduced to the action of compressive or tensile stresses of different magnitude oriented along three mutually perpendicular directions, referred to as the *principal stress axes.*

In an isotropic body the principal stress axes coincide with the principal strain axes. In that case any complex deformation of the

body can be reduced (for an elementary volume) to contraction and elongation along the three mutually perpendicular axes.

In view of that the deformation experienced by the body is sometimes represented as the so-called *strain ellipsoid*. Imagine an isotropic sphere (Fig. 23*a*) representing the initial (non-strained) state of the body. If tensile or compressive forces of different magnitude are applied to the sphere along the three perpendicular axes, the sphere will turn into a three-axial ellipsoid. The dimensions of the axes of the ellipse will represent the relative magnitudes of deformation along the principal axes.

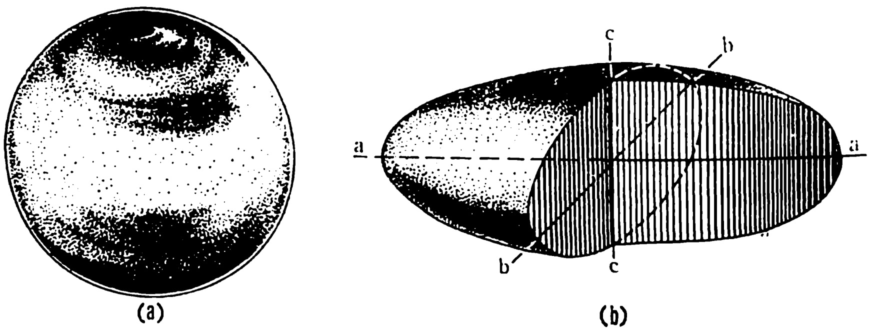

Fig. 23. Deformation represented by the strain ellipsoid:
a—original sphere; *b*—strain ellipsoid; *a-a*—long axis of the ellipsoid (axis of greatest elongation); *b-b*—median axis of the ellipsoid and deformation; *c-c*—short axis of the ellipsoid (axis of greatest contraction); *a-a*, *b-b*, and *c-c*—principal strain axes

Strain is said to be uniform if all the sections of the body are uniformly strained and also if the principal strain axes everywhere retain their directions; it is referred to as non-uniform if the magnitude and the character of strains change from one section of the body to another and if the directions of the principal axes change as well. Non-uniform strain is represented by bending and torsion while uniform strain is expressed in shear and compression-extension in isotropic bodies.

As stresses increase up to a certain limit, a solid body, as a rule, suffers *elastic deformation*. Such deformation is reversible. When the forces are removed the original form of the body is restored. With a further increase of the deforming forces, elastic deformation is replaced by *plastic, that is residual deformation*. Such deformation does not disappear when the forces are removed. When the forces are increased still further, a moment comes when the body is ruptured due to the development of a crack or of a number of cracks within it.

The total stress acting on a given unit area may be resolved into two components: the stress acting normal to the unit area (normal

stress) and the stress directed parallel to the plane of the unit area at some angle (tangential or shearing stress) (Fig. 24).

At a given direction of the external deforming forces, the relationship between normal and tangential stresses will differ depending on the orientation of the unit area in question. For instance, if the body is subjected to extension (see Fig. 24) normal stresses will be greatest on the unit areas normal to the axis of tension, whereas tangential stresses will be absent. On unit areas parallel to this axis both normal and tangential stresses are absent, and on the unit areas of intermediate orientation both normal and tangential stresses are present but in different relationship.

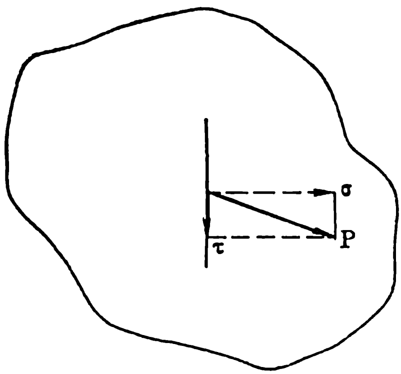

Fig. 24. General (P), normal (σ) and tangential (τ) stresses

As an example, we shall consider a body subjected to tensile stress along one axis (uniaxial tensile stress). Let us designate as σ the normal stress on any unit area, as σ_1 the normal stress on the cross-section of the body (normal to the tensile and compressive forces), as τ the tangential stress on any unit area and as α the angle between the given unit area

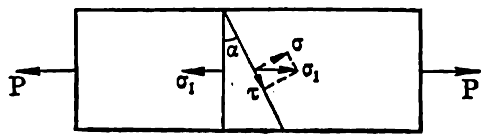

Fig. 25. Diagram of normal and tangential stresses in uniaxial tension

and the cross-section (Fig. 25). Then the value of normal stress on any unit area is given by the formula:

$$\sigma = \frac{\sigma_1}{2}(1 + \cos 2\alpha),$$

and the value of tangential stress will be:

$$\tau = -\frac{\sigma_1}{2}\sin 2\alpha.$$

From these formulas it follows that maximum normal stress σ_{max} develops when $\alpha = 0$, i.e., in the case of uniaxial extension the

normal stress will be maximum on the cross-section, or $\sigma_{max} = \sigma_1$. Tangential stresses on the same unit area equal zero.

It follows from the same formulas that maximum tangential stress will be observed when $\alpha = 45°$, i.e., is confined to planes which form an angle of 45° with the tensile axis. In the case of uniaxial extension maximum tangential stress will be in the following relationship to maximum normal stress:

$$\tau_{max} = \frac{\sigma_{max}}{2}.$$

In the case of uniaxial compression the position of planes under maximum normal and tangential stresses will be the same, but the sign of normal stresses in the formulas will be reversed.

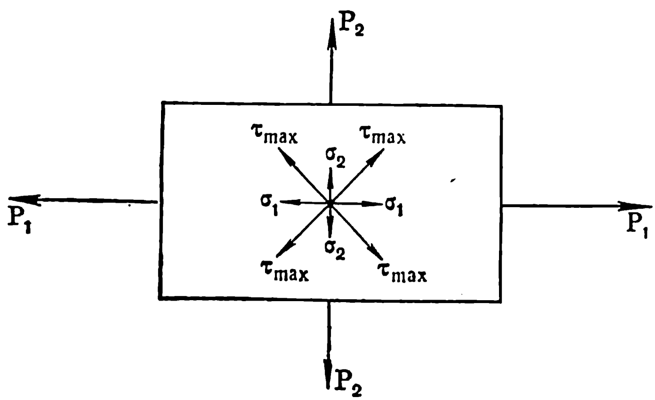

Fig. 26. Distribution of principal and maximum tangential stresses in the case of biaxial tension

If a body is subjected to compression or extension along two perpendicular axes, this is a case of biaxial or plane deformation (Fig. 26). In this case the tensile stresses should be regarded as positive and the compressive stresses as negative. If the forces acting along the two axes have the same sign, one may speak of biaxial strain of the same sign; otherwise it must be referred to as biaxial strain of different signs. A graphic representation of positive biaxial strain of the same sign is given in Fig. 26. Let us designate the strains caused by biaxial stresses along the two axes as σ_1 and σ_2, respectively, so that $\sigma_1 > \sigma_2$. Maximum normal stress will then be observed on planes perpendicular to the direction of the principal tensile stress, and will equal α_1. Maximum tangential stress will be observed on planes that bisect the angles between the planes of the principal normal stresses, that is on planes that form an angle

of 45° with the principal axes. The value of maximum tangential stress in the case of biaxial extension is determined as follows:

$$\tau_{max} = \frac{\sigma_1 - \sigma_2}{2}.$$

If compressive stresses act along one or both axes the signs of the corresponding symbols in this formula should be reversed. If the tensile (or compressive) stresses are equal the numerator in the last formula and τ_{max} become zero. This means that under the given conditions no shearing stresses develop at all. Tangential stresses can develop only if the stresses applied along two perpendicular directions differ among themselves either in sign or in magnitude.

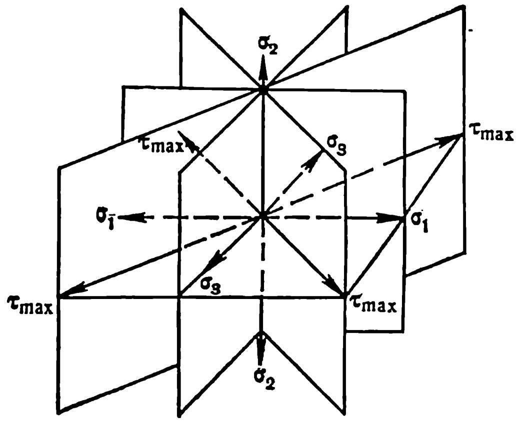

Fig. 27. The principal and maximum tangential stresses in the case of triaxial tension

A diagram of triaxial strain is given in Fig. 27. The three tensile stresses acting along three mutually perpendicular directions are designated as σ_1, σ_2 and σ_3 and furthermore $\sigma_1 > \sigma_2 > \sigma_3$. In this case maximum normal stresses obviously will be observed on surfaces located at right angles to maximum tensile stresses and will equal σ_1. Maximum tangential stress will be confined to two mutually perpendicular surfaces that bisect the right angles between the surfaces of maximum and minimum values of normal tensile stresses. The value of maximum tangential stresses will be:

$$\tau = \frac{\sigma_1 - \sigma_3}{2}.$$

The tangential stresses on surfaces that bisect other angles between surfaces with tensile stresses will be smaller and will be expressed

respectively as

$$\tau_{1,2}=\frac{\sigma_1-\sigma_2}{2},$$

$$\tau_{2,3}=\frac{\sigma_2-\sigma_3}{2}.$$

As in the previous case of biaxial stress, tangential stresses develop only if there is a difference in the magnitude or the sign of the applied tensile or compressive stresses. If a body is under uniform confining stress no tangential stresses develop.

In all the cases discussed above, the basic assumption was that we are dealing with uniform strain in an isotropic body. Geology always deals with anisotropic bodies and non-uniform strain. The distribution of the principal stresses is always complicated and varies from place to place both in magnitude and direction. Furthermore, the distribution of stresses changes with time and so do the mechanical properties of rocks subjected to great plastic deformations.

This considerably complicates the mechanical interpretation of rock deformations observed in nature. The above reasoning applies only to separate small volumes of rocks, within which strain may be regarded as uniform. For complete mechanical analysis it would be necessary to determine the position of the principal stress axes for each small volume of rock separately and take into account the changes in the position of these axes in time.

The entire complex of stresses existing in a body suffering deformation is called the *field of stress.*

The methods of determining the principal stress axes under geological conditions will be considered below when we shall deal with internal dislocations of rocks.

It was noted earlier that as the stresses applied to a body increase, the body first reacts to this action by elastic deformation. As known, this deformation obeys Hooke's law of proportionality between the magnitude of stress and the magnitude of strain:

$$\sigma=E\varepsilon,$$

where σ = stress and ε = deformation.

The proportionality factor E is called Young's modulus.

In the general case, with increasing applied stresses, elastic deformation passes into plastic. The point at which this change occurs is called the elastic limit. An ideally plastic body is characterized by the fact that it does not offer additional resistance to plastic deformation which develops infinitely at the value of stress corresponding to the elastic limit. Actually, plastic deformation of real bodies requires that the stresses exceed the elastic limit. This is

associated with changes in the internal structure of the body in the process of deformation, i.e., with a phenomenon known as strain hardening. Usually, however, following a certain increase in stresses required to initiate plastic deformation, the value of the stresses needed to maintain plastic deformation sharply diminishes. The body, as it were, "yields" and deformation proceeds increasingly. The end result of the process is rupture of the body.

The substance of the transition from elastic to plastic deformation is that the elastic deformation caused by a particular load gradually becomes permanent as a result of rearrangement of particles (atoms, molecules, etc.) which assume a new balanced arrangement which corresponds better to the new form of the body.

This rearrangement of particles is predetermined by their displacement in the process of elastic deformation and their thermal movements in the course of which the particles make random jumps and from time to time find themselves in places more favourable at the given stress distribution and stay there. The stresses that had developed in the process of elastic deformation are gradually dissipated as it were. Such dissipation of elastic stresses is referred to as *relaxation*.

Sometimes relaxation develops rapidly, sometimes slowly which depends on the properties of the body. If relaxation is relatively slow, which means that plastic deformation meets considerable resistance, the body is said to be very viscous. In the case of bodies of low viscosity relaxation and fixation of deformation take place rapidly. If the stresses are maintained, elastic deformation increases parallel with the development of relaxation, as a result of which elastic deformation passes into plastic. Thus, the development of plastic deformation is always accompanied by elastic deformation which at varying rate passes into plastic deformation.

Relaxation is associated with what is known as *creep*—the deformation of a body under prolonged exposure to stress. The substance of this phenomenon is that when a body is subjected to stresses that do not reach the elastic limit but are maintained long enough (provided the rate of relaxation is not too slow), elastic deformation will pass into plastic. This occurs as a result of rearrangement of particles in the course of thermal movement. The stresses that existed prior to that will dissipate and ever smaller stresses will suffice to sustain deformation of a certain magnitude.

Theoretically, any load, however small, can cause plastic deformation of any body provided enough time is allowed for the deformation to develop. In a number of cases, however, the time required for the development of observable deformations may be too long even on the geological time-scale. It is assumed that real solid bodies, including rocks, have a certain "yield point", that is to say that

below a certain minimum value of stress plastic deformation does not develop, no matter how much time is allowed for it.

Relaxation and creep play a most important part in geological processes making possible the development of major plastic deformation in the earth's crust under the action of stresses that are relatively not great, but act for millions of years.

The internal mechanism of plastic deformation is associated with shearing stresses and relative displacement of the particles of a body along the planes of maximum shearing stresses. For example, plastic elongation of a wire occurs by relative sliding of very thin plates

Fig. 28. Displacements in a wire in plastic elongation

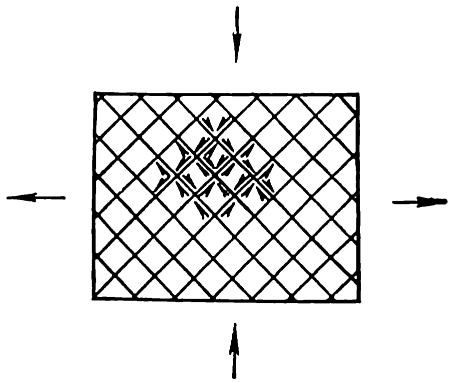

Fig. 29. A diagram of slip planes within a body in the process of plastic deformation. Small arrows indicate the directions of slip along the slip planes

along planes inclined at an angle of 45° to the axis of tension (Fig. 28). The existence of conjugate planes of maximum shearing stress results in a complex picture of gliding within the body in the course of deformation. Fig. 29 shows diagrammatically the slip planes in a body subjected to compression in one direction and to tension in another. The slip planes divide the body into a multitude of minute prisms, some of which are being squeezed out and others wedged in between.

Polycrystalline and particularly granular structure of rocks introduces specific features into the plastic deformation of rocks. A granular rock usually has planes of weakened cohesion between the grains. The slip developing in the course of plastic deformation can exploit these weak planes. Such may be, for example, the contact planes between individual grains. In that case the slip does not penetrate inside the grains and they are not deformed. However, they do move relatively to one another, are turned around and rearranged as a result of their displacement, owing to which the rock

as a whole suffers deformation. This intragranular slip is typical of loose uncompacted rocks.

In harder rocks, in which intergranular cohesion is stronger, slip may develop more easily not along intergranular contacts but in the grains themselves. In this case intragranular plastic deformation occurs in the process of intragranular slip and results in the deformation of not only the rock as a whole but also of each grain separately.

The deformability of a body depends on its mechanical properties, which in turn depend on external factors, such as temperature, solvents and adsorbed liquids, confining pressure, rate of deformation, the direction of external stresses.

An increase in temperature contributes to the plasticity of solid bodies; they become softer and are more easily deformed. Contact of a body with a solvent or an adsorbed liquid also increases its plastic deformability.

Confining pressure has a twofold effect. On the one hand, it increases the resistance of a body to deformation which means that greater stresses are required to produce the same deformation. On the other hand, the same factor greatly increases the plasticity of bodies in the sense that they can withstand great plastic deformation without rupture. Limestone, for example, at atmospheric pressure suffers only elastic deformation and breaks immediately beyond the elastic limit without the slightest plastic deformation. At a confining pressure of 10,000 atmospheres a bar of the same limestone may be subjected to plastic elongation amounting to 50 per cent of its initial length without rupturing.

The effect of the rate of deformation on the properties of bodies is that as it increases their resistance to deformation is increased and plasticity descreased. Conversely, a slower deformation rate increases the plasticity of bodies. This is the phenomenon of creep mentioned earlier. Small forces slowly acting on a body can gradually cause great plastic deformation, whereas when the forces are rapidly applied they must be much greater to produce deformation of the same magnitude. Furthermore, in the case of rapid action plastic deformation may not occur at all because the body will rupture before it does.

The direction of external stresses also affects the properties of a body. The point is that plastic deformations are more readily caused by compressive stresses than by tensile stresses; a body subjected to tension always proves to be more brittle than one under compressive stresses and may fail before great plastic deformation has developed.

In geological conditions the time factor is most important. It is this factor that apparently determines the ability of ordinarily

brittle rocks such as sandstones or limestones to bend into complex folds without rupturing, as if they were as soft as clay. Possibly, increase of temperature with depth also plays a part. Moisture present in rocks is another factor increasing their deformability.

7. THE MECHANICS OF ROCK FAILURE

It is known that any deformation ends with rupture of a body, provided the stresses reach a value corresponding to the ultimate strength of the given body.

Two kinds of rupture are known: *tensile fractures* and *shear fractures*. Tensile fracture is caused by normal tensile stresses. It develops when these stresses attain a certain critical value and is expressed in the development of a fracture perpendicular to the principal strain axis (Fig. 30).

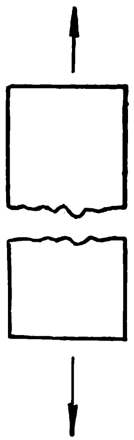

Fig. 30. Development of rupture under tensile stress

Shear is caused by shearing stresses and is expressed in the development of fractures oriented in conformity with the direction of maximum shearing stresses. Maximum shearing stresses are observed on planes which lie at angles of 45° to the extension-compression axis. However, it should be mentioned that actually shear fractures rarely coincide with the theoretical direction of maximum shearing stresses. In real solid bodies, including rocks, the angle between shear fractures and the axis of principal compressive stresses is always less than 45° (from 30 to 45°). This deviation of shear fractures from the theoretical direction bears relation to the composition and internal structure of bodies and is apparently determined by friction forces. These develop along the fracture plane, inasmuch as in shear, the shearing stresses cause a certain, though very small displacement of the walls of fractures relatively to one another. The actual position of the shear fracture is determined by the most favourable combination of values of shearing stresses and friction along the fracture plane (Fig. 31). Hereafter, however, for simplicity we shall assume that shear fractures develop along maximum shearing stress planes.

Inasmuch as plastic deformation is caused not by normal stresses but by shearing stresses, rupture is not directly connected with plastic deformation. It may, and often does, occur right after elastic deformation, which is known as brittle failure. In this case the critical point is achieved, while elastic deformation is still in pro-

gress. It has been established, however, that tensile strength depends on the rate of deformation and therefore on the time of action of forces on a body. In rapid deformation greater stress is needed for brittle failure to develop.

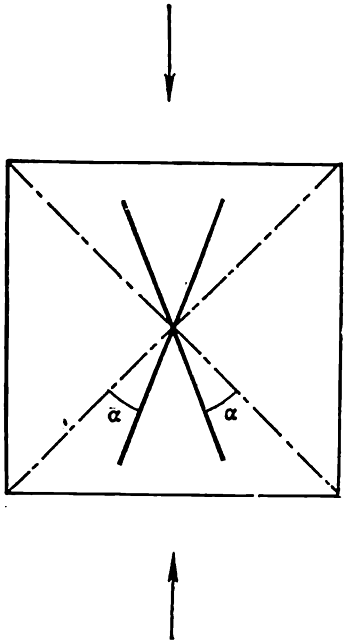

Fig. 31. Shear fractures developing in compression:

dashed lines—theoretical position of fractures; solid lines—actual position of fractures which deviated from the theoretical by an angle α

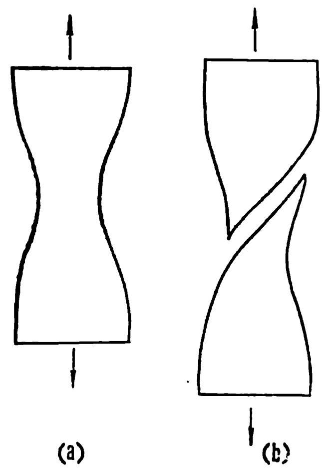

Fig. 32. Viscous shear upon tension:

a—development of a neck; b—development of a shear fracture

Rupturing does not occur all at once. First there appear small incipient cracks which gradually merge into a single fracture. Tension fractures typically have uneven jagged sides.

Shear too may be brittle and develop right after elastic deformation. Most often however, shear fractures develop after plastic deformation. In these cases shear is closely related to preceding plastic deformation, and constitutes, as it were, its further continuation.

It is known that plastic deformation consists of slipping of plates within the body that is suffering deformation in the direction of the maximum shearing stress. A certain period of strain hardening is followed by weakening of the body and, though the acting force

remains the same, plastic deformation proceeds at a quickening pace. This is due to the development of the so-called neck in the test specimen (Fig. 32). Strain that originally was uniformly distributed over the entire specimen, now concentrates in a certain part (the middle part in our drawing) where the specimen suffers much more rapid deformation than in other parts. Gradually the deformation process concentrates in a still more narrow zone until a shear fracture develops in this zone and the specimen splits in two. From the mechanical point of view this process can be conceived as follows: originally the slipping within the body occurs uniformly along numerous planes; further on slip concentrates on fewer and fewer planes, due to which it becomes more rapid on each such plane. Finally, the concentration of slipping, confined to one plane, reaches its limit and the speed of movement along this plane becomes so great that failure develops along the plane.

This is known as viscous failure in contrast to brittle failure.

A viscous shear fracture always develops gradually. First small cracks appear in various parts of the body, these gradually join each other and grow. The ends of the cracks often branch out and in this zone of branching the disjunctive dislocation gradually passes, as it were, into plastic deformation which continues in the adjacent segments of the body in which deformation has not yet resulted in fracture.

The patterns of ruptures and shear fractures developing in bodies subjected to various strains are shown diagrammatically in Fig. 33. As evident from the preceding discussion, the position of the fractures is controlled by the direction of maximum normal tensile stresses and maximum shearing stresses in each specific case. Knowledge of these patterns is essential for interpretation of tectonic fractures in geological practice. The position of fractures given in Fig. 33 is the theoretical one which coincides with the direction of maximum shearing stresses.

At this point mention should be made of another important factor that affects the position of shear fractures and changes the angle between the observed fracture and the axis of maximum compression. The point is that in geological conditions viscous fractures develop and gradually grow concurrently with continuing plastic deformation in the surrounding rocks. In the process of this plastic deformation, the earlier-developed fractures may be displaced, turned or bent; as a result their position relatively to the strain axes may change considerably.

Special note should be made of those systems of fractures that develop under the conditions of shear strain, i.e., under the action of a couple of forces. Fig. 34*a* shows the shear of a square body caused by forces that acted in the upper part of the body from left to right

and from right to left in the lower part. The dashed arrows along the vertical sides show the "reactive" forces, the existence of which must be inferred to explain why the body which is subjected to strain does

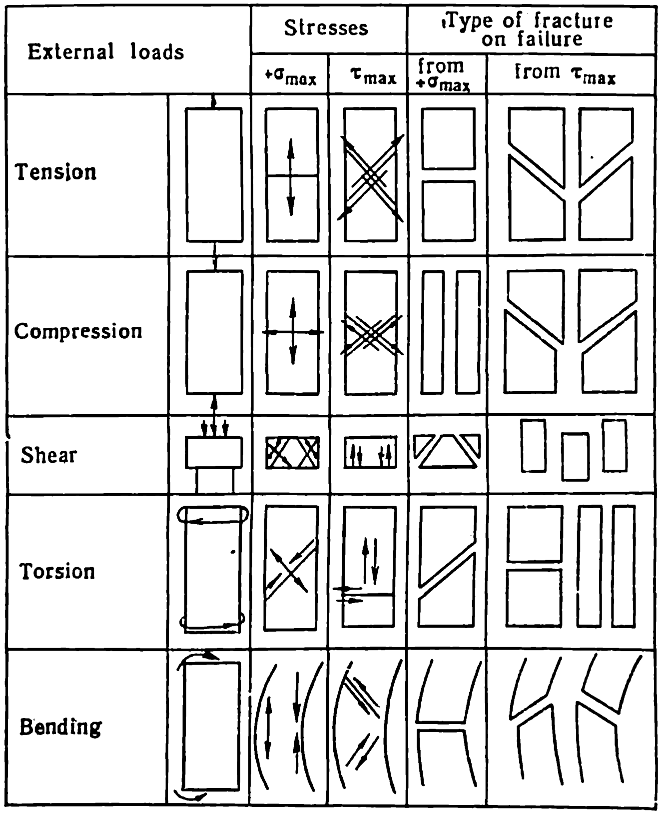

Fig. 33. Diagrammatic representation of ruptures and shear fractures under different kinds of stresses (after Y. B. Fridman)

not turn around. The arrowheads within the figure show the direction of the axes of maximum compression ($-\sigma_{max}$) and of maximum tension ($+\sigma_{max}$).

Fig. 34*b* shows the direction of maximum shearing stresses in the same body. One system of these stresses is parallel to the couple of acting forces and the other is at right angles to it. Shear fractures should develop in these directions. As to tension fractures they must develop at right angles to the axis of maximum tension, i.e., as

evident from Fig. 34*a* parallel to the diagonal of the square which connects its upper left corner with the lower right corner. This direction is shown in Fig. 34*b* by the solid line. If a separate small square is replaced by a whole band of rocks subjected to shear, it would be clear that numerous tension fractures may develop within such a band. They will lie parallel to each other but at an angle of 45° to the axis of the zone subjected to deformation (Figs. 35 and 36). Such a system of tension fractures is known as *echeloned fractures*. Their orientation is controlled by the direction of the principal

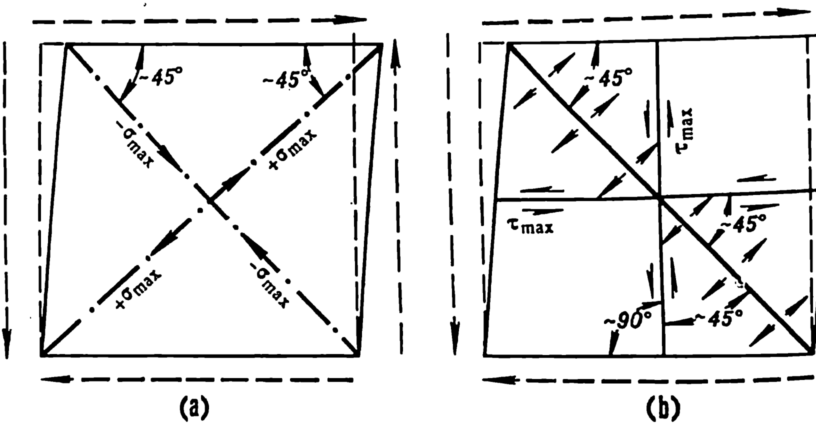

Fig. 34. Development of shear fractures (stress diagram):
a—position of principal axes of tension ($+\sigma_{max}$) and compression ($-\sigma_{max}$); *b*—direction of maximum shear stresses (τ_{max}) and direction of tension fractures (thick diagonal). Thick dashed lines with arrowheads—couple of external active forces; thin dashed lines with arrowheads—couple of reaction forces that restrain the body from turning. Small arrowheads within the diagram—direction of stresses and displacements

couple of forces. If these are oriented in a direction opposite to their orientation on Fig. 34 (that is to the left at the top and to the right at the bottom) the tension fractures too will be differently oriented (in the plane of the drawing they will be inclined to the left and downward). Therefore, the direction of tension fractures within the echeloned system relatively to the general direction of the system can be used to ascertain the direction of shear deformation. If plastic shear deformation continues in the adjacent areas, the echeloned fractures keep growing, earlier-formed parts of fractures may swerve while the new parts will grow in the former direction, controlled by the position of the axis of maximum tensile stresses. As a result the fracture becomes sigmoidal (Fig. 37).

This picture of development of shear fractures is a theoretical one which takes no account of deviations of shear fractures in real bodies

from the plane of maximum shear stresses. Therefore it may be concluded that echeloned series can be formed by shear fractures as well.

Under given conditions either shear or rupture is observed. However, when conditions change the type of failure may change too.

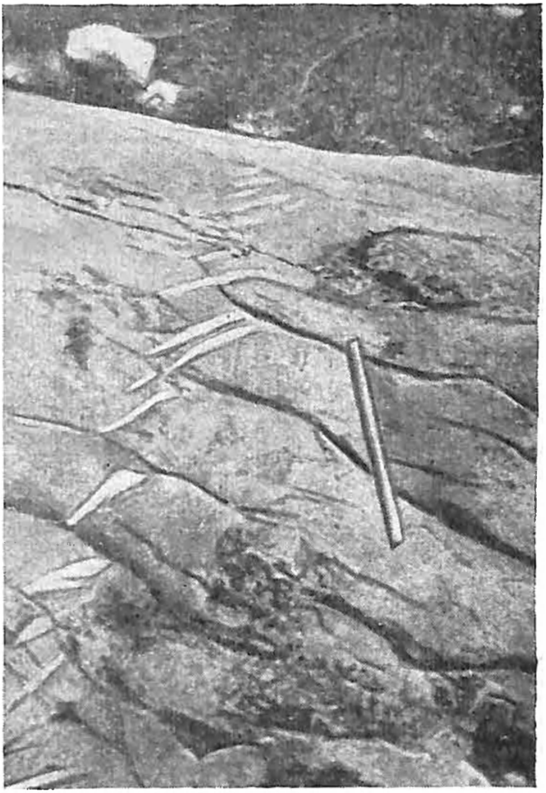

Fig. 35. Echeloned series of tension fractures in nature: the left-hand part of the mass was displaced away from the reader; the right-hand part—towards the reader

Solid bodies are characterized by two kinds of strength, known as shear strength and tensile strength. They are measured by the value of the stresses at which failure occurs. The relative value of either kind of strength depends not only on the properties of the material

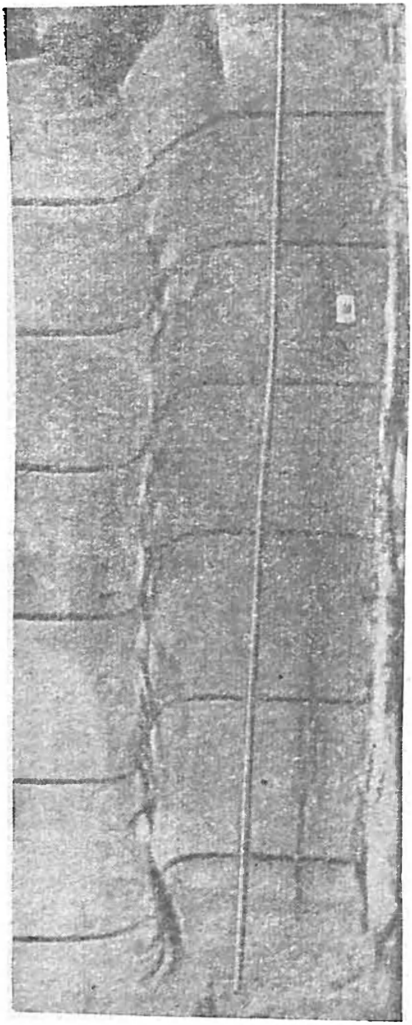

Fig. 36. Echeloned tension fractures obtained on a clay model:
the right-hand part of the model moved upwards; the transverse scratches originally were straight (experiment by M. V. Gzovsky)

but also on external conditions—temperature, confining pressure, rate of deformation, etc. Naturally under given conditions, failure occurs in the way which requires the least stresses.

For example, increasing temperature lowers shear strength but has little effect on tensile strength. Therefore with increasing temperature it is more likely that failure will occur through shear. Conversely, with declining temperature, shear strength increases and tension fracturing becomes more likely.

A higher rate of deformation has the same effect as declining temperature: shear strength increases more than tensile strength. As a result the material becomes more brittle and more prone to rupturing than to viscous shear. Conversely, in the case of slow deformation, failure by viscous shear is more likely.

In nature, the development of fractures, their direction and the order in which they developed are affected to a considerable degree by cavities and various nonhomogeneities in the rocks. Greater stress concentration is observed near cavities, in view of which at appropriate stress values they tend to increase. Earlier-developed fissures affect the stress distribution and, therefore, the position of later fractures. This is associated, for example, with the widely observed arching of fractures. As soon as a small fracture appears, the "force lines" around it are bent, due to which the fracture does not develop in the same direction but deviates from it.

As a result of nonhomogeneities in the composition of rocks fractures tend to concentrate near the weakened boundaries between the

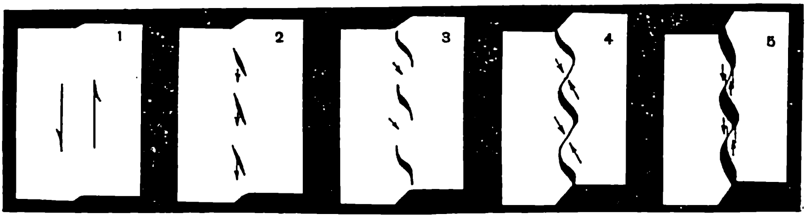

Fig. 37.
Stages 1-3—sigmoid bending of echeloned tension fractures upon shear deformation and simultaneous growth of fractures. Stages 4 and 5—merging of tension fractures into a single sinusoidal rupture (after M. V. Gzovsky)

rocks (contacts of an intrusive mass with country rock, bedding planes and "dead" tectonic faults, etc.) and sometimes deviate considerably from the theoretical position determined by the applied stresses.

CHAPTER III

Folding

The term *folding* covers tectonic dislocations without a break in continuity which are differently manifested in stratified sedimentary, metamorphic and effusive rocks on the one hand and in massive (intrusive) rocks on the other. In the first place, in stratified rocks the results of dislocations are more clearly recorded because due to the form of the strata it is possible to observe the "pattern" of dislocation in different parts of the series. In the absence of stratification the record of deformation is considerably more obscure. Second, the difference is also explained by the fact that stratified rocks suffer deformation in a different way from massive rocks under the same conditions.

Here we shall consider mostly dislocations occurring in stratified rocks. Dislocations occurring in massive rocks will be considered later.

8. MONOCLINES AND FLEXURES

We had said earlier that horizontal strata constitute primary structural forms of sedimentary rocks. If one segment of the earth's crust is uplifted relatively to an adjacent one without a break in continuity of rocks, the strata are tilted. This mode of occurrence in which rocks dip more or less uniformly in one direction only is known as a *monocline*. It is characterized by the strike of the strata within its limits, their dip in one direction and also by its width and length. The term "monocline" is commonly applied to those cases when strata over a great areas have a uniform gentle dip, this occurs mostly on platforms.

Tracing a monocline along the strike and dip one may notice additional bending of strata within it resulting in local alterations of the dip angle of the rocks. The areas characterized by more gentle attitude of strata against the background of the general dip of the monocline are known as *structural terraces*. If such a structural terrace looks like a platform extending in the direction of the dip of the monocline, it will appear on a contour map as a cape. Such a structure is known as a *structural nose* (Fig. 38).

A bend in a monocline characterized by a steeper dip of the strata is called a *flexure*. This term is also used to describe a step-like bending of strata against a background of horizontal bedding (Fig. 39). A flexure consists of an upper limb, an inner and a lower limbs. It may be characterized by the dip angle of the strata within all the

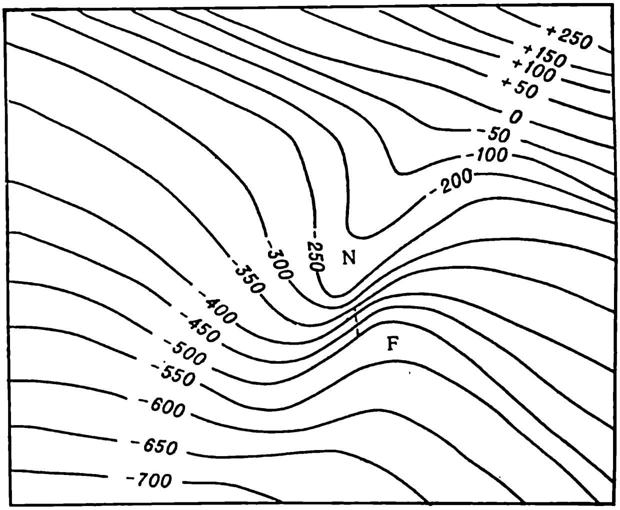

Fig. 38. A monocline complicated by a structural nose (N) and flexure (F). The contour lines are drawn at 50-m intervals

three limbs, the length of the inner limb and by its vertical amplitude. The amplitude of a flexure always varies along the strike and every flexure terminates at a certain distance. The dip angle of the inner limb changes as well.

Flexures often occur in groups, which are then called *ladder flexures*.

Sometimes the term "horizontal flexures" is used to describe the bending of overturned strata observed on the horizontal section.

9. UPWARPS AND DOWNWARPS

If adjacent segments of the earth's crust have experienced relative uplifts and subsidences without a break in the continuity of the strata, these will exhibit corresponding upwarps and downwarps. When such upwarps and downwarps occupy vast areas and dip at a very low angle they are referred to as *anteclises* (upwarps) and *syneclises* (downwarps).

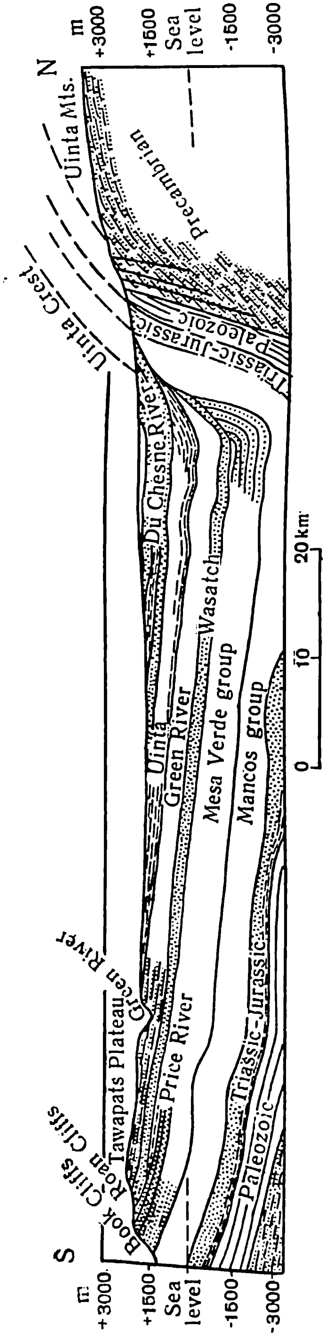

Fig. 39. A flexure in the Rocky Mountains (USA). The flexure is formed by Triassic-Jurassic and younger deposits; in more ancient strata the flexure is replaced by faults (after Irdley)

The throw of downwarping of a syneclise or uplifting of an anteclise may reach two or three kilometres, but they may be hundreds and even thousands of kilometres wide. The gradient of the strata on the limbs usually amounts to a few metres per kilometre, which corresponds to a few minutes of the angle of inclination.

Anteclises and syneclises are typically characterized by a steeper attitude of deep strata and relatively gentler dip of the upper strata. This variation in the attitude of the strata from the bottom upward goes hand in hand with regular variations in thickness. The strata are thicker in the central part of the syneclise and thinner on the limbs and arch (Fig. 40). Many series present in syneclises often completely pinch out in anteclises.

Syneclises and anteclises are typical of platforms. They characterize the mode of occurrence of rocks which compose the upper structural stage of the platform as such. It is known that the structural platform stage is always underlain unconformably by strongly dislocated and more or less metamorphosed basement rocks. The gently bent and eroded surface of the latter serves at the same time as the foundation for the platform series that form the syneclises and anteclises.

Those anteclises at whose crests the folded foundation outcrops on a large area because the sedimentary cover has been eroded, are known as *shields*. An example is the great Baltic Shield comprising Karelia, the Kola Peninsula, Finland and Sweden, within which

strongly dislocated and metamorphosed Precambrian rocks are exposed.

An example of a syneclise is the Moscow Basin (of late it is often referred to as the Moscow syneclise). It is known that in the central part of this syneclise near Moscow the crystalline basement lies at a depth of about 1,600 m. This basement outcrops to the surface in two adjacent anteclises—the Baltic (the Baltic Shield) in the north, and the Voronezh anteclise in the south. The crystalline basement is covered by sedimentary rocks of the Paleozoic and Mesozoic times

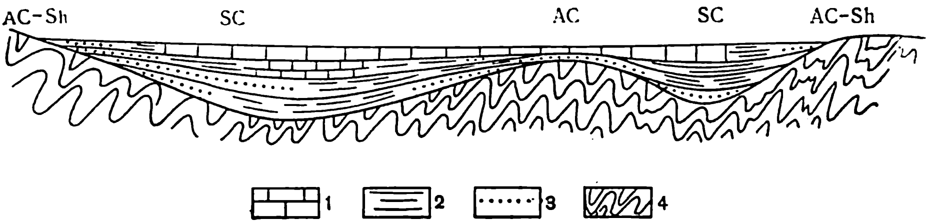

Fig. 40. Schematic geological profile across syneclises (SC), an anteclise (AC) and shields (AC-Sh):

1—limestones; *2*—clays; *3*—sands; *4*—folded basement. The vertical scale is strongly exaggerated.

whose shape generally resembles a giant bowl. From north to south the syneclise is 1,300 km across. Thus, at the slopes of the syneclise the gradient of the strata amounts approximately to 2.5 m per kilometre. The sedimentary section is thickest and most full in the central part of the syneclise, where it is composed of Lower Paleozoic, Devonian, Carboniferous, Jurassic and Cretaceous rocks. Towards the margins of the syneclise all these deposits gradually pinch out.

By analogy with limbs of folds, the slopes of syneclises and anteclises are also referred to as "limbs" (see below). The most convex part of an anteclise may be called its crest. The limbs of syneclises and anteclises are often complicated by flexures and structural terraces.

More sharp structural forms called anticlinal uplifts and tectonic depressions also belong to the group of upwarps and downwarps. They differ from anteclises and syneclises mostly in shape and size. Anticlinal uplifts and tectonic depressions appear on plan as elongated oval upwarps and downwarps, as different from typical anteclises which are round or irregular. The length of individual anticlinal uplifts and tectonic depressions is usually measured in hundreds of kilometres and the width in terms of a few tens of kilometres. The amplitude of an anticlinal uplift or tectonic depression amounts

to a few kilometres. The average slope of strata on the limbs of these structural forms is not great (5 to 10°) but parts of the slopes complicated by flexures may dip at a much higher (up to 30° and more) angle. Anticlinal uplifts and tectonic depressions are often asymmetric, with limbs of different steepness.

Anticlinal uplifts and tectonic depressions rise and plunge along the strike, being divided, as it were, into separate oval uplifts and sags. In either direction along the strike they terminate periclinally and centroclinally like very large brachy-folds, of which more below.

Anticlinal uplifts and tectonic depressions may branch along the strike into two or more structural forms of the same type. This is known as *virgation*.

Anticlinal uplifts may form isolated upwarpings within regions characterized by a generally flat bedding of strata. More often, however, they occur in parallel groups, alternating with depressions. The latter have no importance of their own and are always associated with the uplifts, being located on either side of each uplift and compensating, as it were, by their subsidence the elevation of the uplifts.

A typical isolated anticlinal uplift is the Karatau Ridge on the Mangyshlak Peninsula. Judging by the attitude of the Jurassic, Cretaceous and Paleogene strata of which this ridge is composed it may be regarded as an elongated oval-shaped upwarp of the earth's crust extending from southeast to northwest for 200 km and about 40 km wide. On the limbs the strata slope at an angle of 3 to 6° with local increases to 35°. The upwarp is flanked by depressions of similar dimensions filled with Tertiary strata.

A country of anticlinal uplifts and tectonic depressions is the Tien Shan range. In northern and southern Tien Shan the uplifts and depressions are formed by Mesozoic and Cenozoic strata. A typical anticlinal uplift is the Ketmen ridge of northern Tien Shan. This is an oval uplift (Carboniferous strata overlain by Jurassic outcrop to the surface), extending longitudinally for 90 km and up to 35 km across. The slope of the strata varies from 5 to 30°. Judging by the attitude of Mesozoic and Cenozoic strata, such great ridges as the Trans-Ili Alatau, the Kirghizian range and the Turkestan or Alai mountains are also anticlinal uplifts. By now, however, these great uplifts, which are hundreds of kilometres long and tens of kilometres wide, are strongly disturbed by various dislocations (faults) which distorted their original character.

A typical tectonic depression is the basin of Lake Issyk-Kul filled chiefly with Tertiary sediments which are 4 to 5 km thick in its central part. The oval-shaped depression extends longitudinally for 200 km and is about 70 km wide.

Intermediate structural forms link the anticlinal uplifts and tectonic depressions with typical anteclises and syneclises on the one hand, and (as we pass to smaller uplifts and depressions) with folds of the intermittent type (of which more below) on the other.

Therefore, in a number of cases it is difficult to draw a clear line between these structural forms, and it is quite likely that what one researcher regards as an anticlinal uplift, may be regarded as a big brachyanticline by another. Attempting to classify natural phenomena we invariably observe gradual transitions from one form to another. This, however, cannot compromise the principle of classification. The different structural forms recognized by us are quite specific and can be positively identified, though various transitional forms may require certain qualifications. It should be noted that in literature anticlinal uplifts and tectonic depressions are sometimes called *meganticlines* and *megasynclines.*

Upwarps and downwarps develop on young platforms and in regions of post-platform activization.

They are produced by slow differential uplifting and subsidence of segments of the earth's crust. Crustal movements of this kind are referred to as oscillatory tectonic movements. Undoubtedly the earth's crust throughout its thickness is involved in these movements. In view of that the structural forms considered in this section are referred to by some researchers as crustal or abyssal folds, and also as *primary* folds in contrast to secondary, derivative and superficial folds which involve only certain upper sequences of the earth's crust. Upwarps and downwarps constitute the structural background for the development of folding described below.

10. THE PRINCIPAL MORPHOLOGICAL TYPES OF FOLDS

Bends of strata, that are more sharply pronounced than syneclises and anteclises and smaller than anticlinal uplifts and tectonic depressions are called *folds*. They vary greatly both in size and shape. The specific features of the shape of a fold are differently expressed, depending on whether we view a particular fold in the vertical section or in plan. On the basis of these morphological differences between folds it is possible to identify among them a great number of varieties, many of which undoubtedly differ in origin as well. In this section we shall consider only the morphological features of different folds and will try to systematize them.

First, we shall examine the features expressed when folds are viewed in the vertical cross-sections.

Anticlinal and synclinal folds, also called anticlines and synclines are recognized. In the first case, as a rule, we deal with a convex fold on whose slopes the strata dip in opposite directions; in the

second case the folds are mostly concave and the strata usually dip towards each other. We say "as a rule" and "usually" because, as it will be evident from the following discussion, in anticlines and synclines the strata may also dip in the same direction and, furthermore, anticlinal and synclinal folds may not be (as an exception) convex or concave respectively.

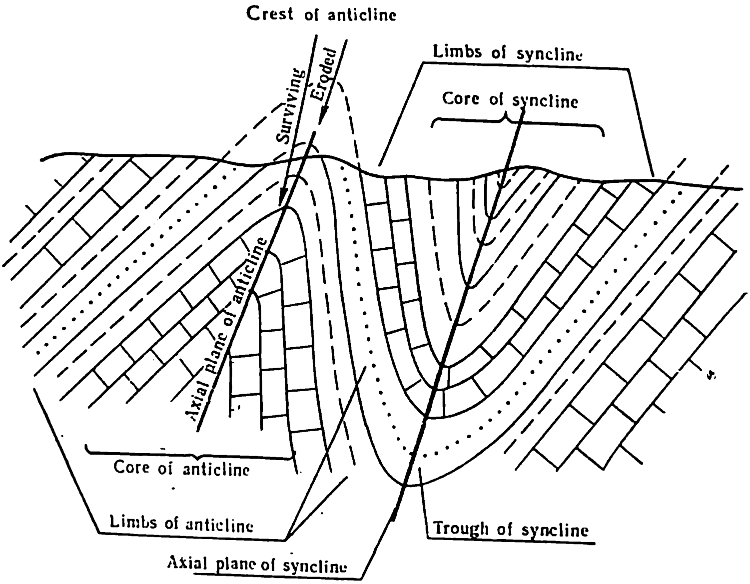

Fig. 41. Elements of folds: core, limbs, curves, and axial planes of an anticline and a syncline

Every fold, anticlinal or synclinal, consists of a *core* and *limbs* (Fig. 41). The core, to use the most general definition, is composed of rocks occurring in the inner part of the folds, while the limbs are composed of rocks occurring along both sides of the fold. These concepts are largely conventional, since, depending on the depth of the section along which the fold is examined, the core and the limbs may be composed of rocks of different ages. Similarly, the relationship between the thickness of the core and of the limbs may be arbitrary. These concepts, however, are very convenient from the practical point of view, since they facilitate description of folds. Sometimes the distinction between the core and the limbs can be more positive, for example, when the inner part of a fold and its sides are composed of different rocks.

The point where the strata curve from one limb to the other is called the *crest* of the fold (see Fig. 41). The crest of a fold may be

considered as the opposite of the limbs, as the zone separating the two limbs; the crest of an anticlinal fold is also called the arch. The corresponding term for the curve of a synclinal fold is trough.

In the cross-section of every stratum in the crest of the fold one can conceive a point at which the stratum bends from one limb to the other.

Visualizing a countless multitude of such points situated along the crest of a fold and connecting them with a line we obtain the so-called hinge of the fold. The hinge of an anticline is the line that connects the highest points of the crest along any one stratum. In a syncline the hinge will be the line which connects the lowest points of the trough along any one stratum (Fig. 42).

If all such hinges within a fold are plotted in such a manner that they touch one another, they will give the *axial plane* or *surface of the fold*. On geological cross-sections it is only possible to plot the trace of this plane with the vertical plane. The trace of the axial plane with the surface of the earth (or with the horizontal plane) is the axis of the fold*. The difference between the axis of a fold and its hinge is that the latter always remains within a single stratum whereas the axis of a fold plotted in the horizontal plane or on the earth's surface may pass from one stratum to another. This may be due either to irregularities of the surface relief, or (if the apical line is drawn in the horizontal plane) to the fact that a more or less pronounced undulation of the strata is always observed along the strike of the fold resulting in transverse anticlinal or synclinal bends (Fig. 42).

In cross-section, folds differ from one another in the relationship between the limbs and the crest. Thus there are recognized *sharp* folds, folds with a rounded crest, *open* folds (the limbs dipping away from the crest), *isoclinal* folds (the limbs along the greater part of the fold are parallel), and *fan-shaped* folds. The latter type is characterized by transverse compression of the fold which may result in complete detachment of the core. There are also recognized coffer or box folds which have a broad flat crest and steep limbs. The types of folds mentioned above are shown diagrammatically in Fig. 43.

Folds may also be distinguished by the inclination of the axial plane and the limbs. If the axial plane is vertical and both limbs are almost symmetrically inclined the fold is said to be *upright*. If the axial plane is inclined and the limbs are asymmetrical (one limb is more steep than the other), the fold is said to be *inclined*. If at one limb the position of the strata is reversed this is an *overturned* or *inverted* fold. When the axial plane is horizontal the fold

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* Sometimes what is called here the hinge of a fold is erroneously called its axis.

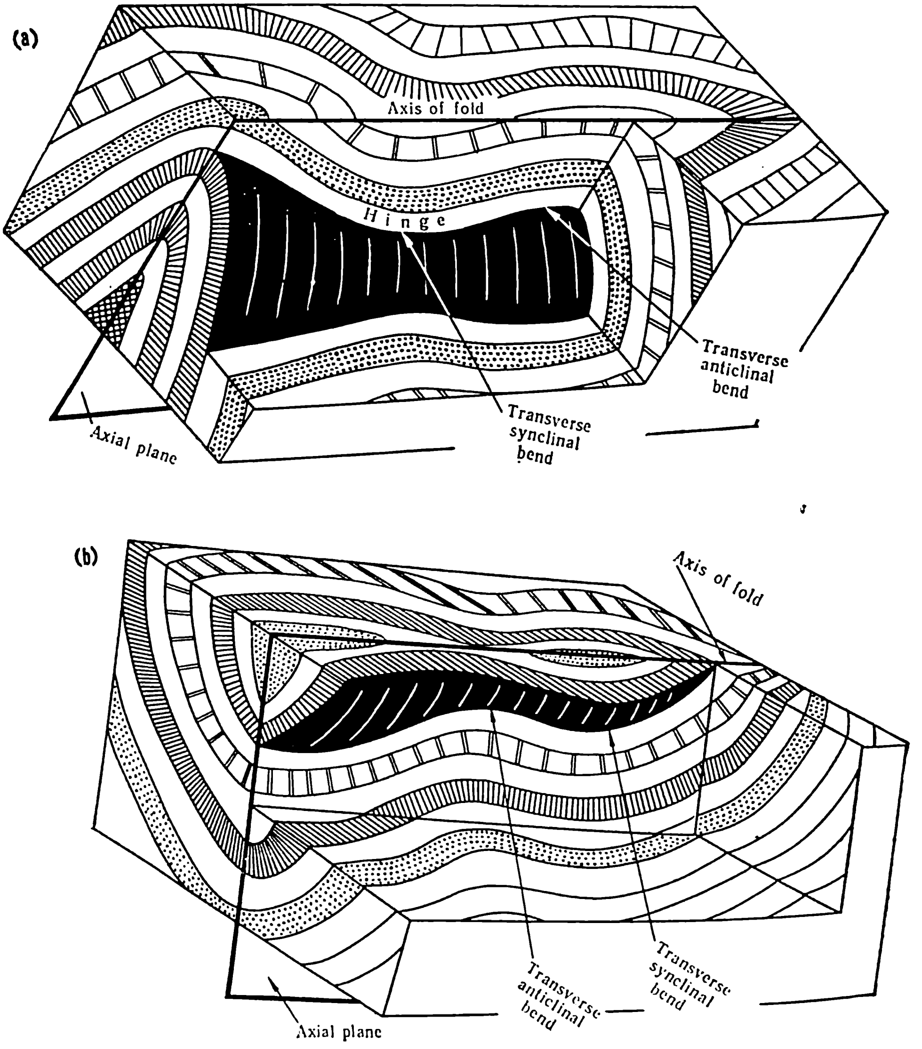


Fig. 42. Elements of a fold:
a—axis and apex of an anticline; *b*—ditto of a syncline

is said to be *recumbent* or *lying*. Furthermore, in some cases the axial plane is depressed below the horizontal position and the fold is described as *dipping*. In that case the arches of anticlinal folds

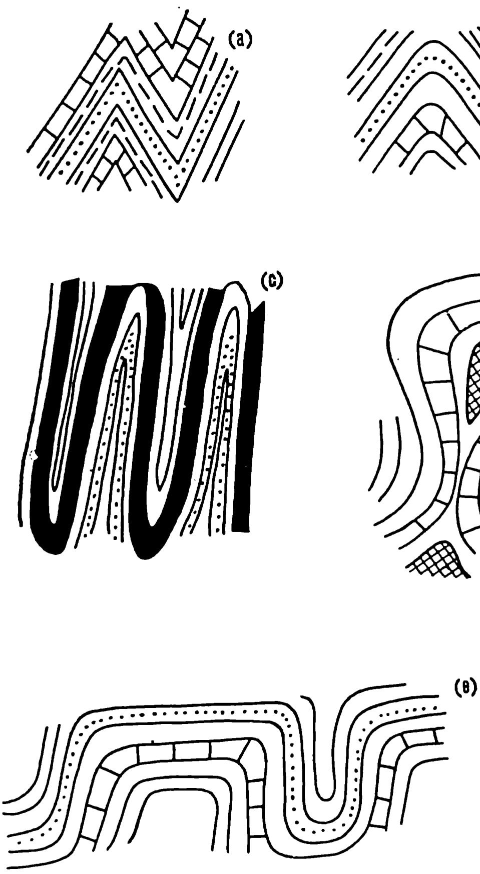


Fig. 43. Folds of different shape:
a—sharp; *b*—rounded open; *c*—isoclinal; *d*—fan-shaped; *e*—box-type

along the bedding of the strata appear as synclines, that is become concave instead of being convex. Conversely in such cases synclinal folds appear as anticlines.

Such folds may be called pseudosynclines and pseudoanticlines. They differ from normal folds in the relative position of rocks of different ages in the core and the limbs. In a normal syncline the core is composed of younger rocks, while in a pseudosyncline the older rocks are found in the core (Fig. 44).

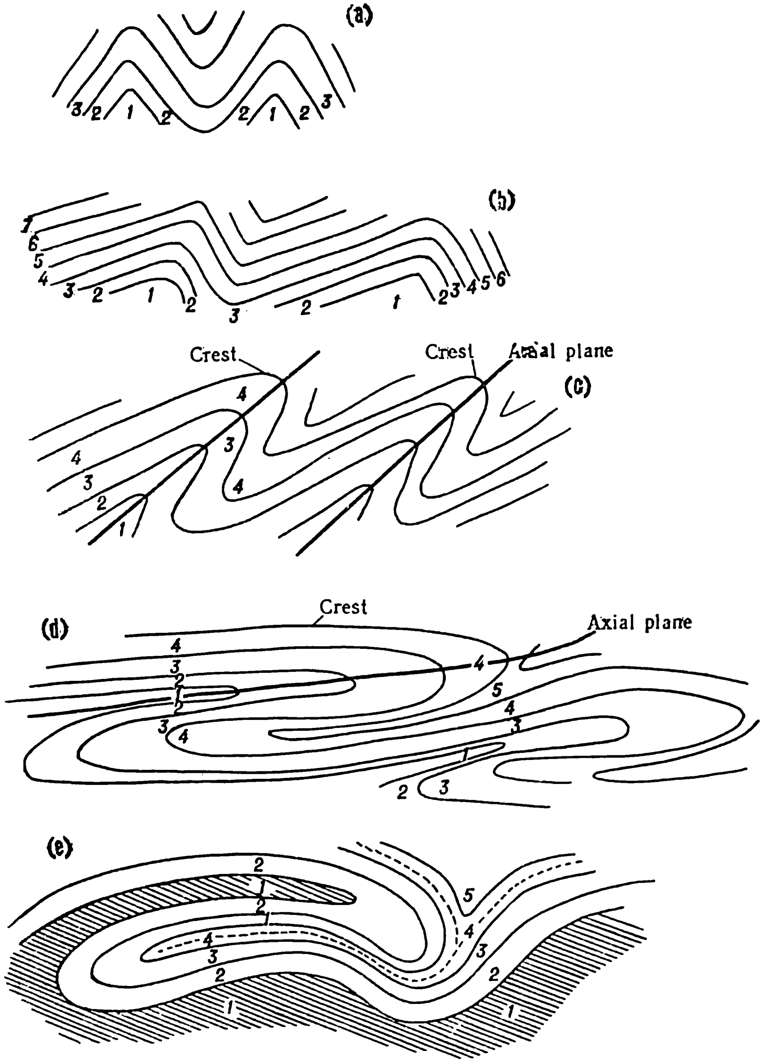


Fig. 44. Folds of different shape (depending on the inclination of the axial plane and limbs):

a—upright; *b*—inclined; *c*—inverted; *d*—recumbent; *e*—dipping (figures from *1* to *6* designate the strata from older to younger)

Considering the occurrence in nature of dipping or overturned folds and pseudoanticlines and pseudosynclines, the above definition of anticlinal and synclinal folds must be qualified. As we have seen, it would be wrong to say that an anticline is always convex and a syncline concave. Therefore, an anticlinal fold must be defined as one whose core is composed of rocks older than those in its limbs. A syncline is a fold whose core is composed of rocks younger than those on the limbs.

This definition covers all cases that occur in nature.

Considering the existence of inclined and recumbent folds it is necessary to mention yet another element of their structure. The point is that in such folds their hinge will not always lie at highest line drawn through a given layer as it does in upright folds. In some cases the highest part of a stratum that forms a part of the fold will occur in the limb. Considering the practical implications of such highest points (and lines) of strata (oil and gas accumulations are usually confined to them), the term *anticlinal apex* has been introduced (see Fig. 44*c*, *d*).

One may also speak of the apical plane of an anticline. A similar terminological problem is presented by the lowest line within a certain stratum of a syncline. There is no commonly accepted term for this case, but sometimes it is referred to as the keel of a syncline.

Considering that folds have many morphological varieties their description is necessarily complex. For example, a fold may be isoclinal and at the same time recumbent, or it may be open with a rounded arch and simultaneously inclined, etc.

Speaking of the shape of a fold we primarily have in view how it is expressed in the attitude of a relatively thin series of strata. Practically always the general appearance of folds changes vertically. At different elevations the same fold may have different features. For example, an upright fold may turn into an inverted one in higher strata, etc.

The configuration of folds in plan exhibits considerable differences (Fig. 45), primarily in the behaviour of the axes of folds along the strike. It has been noted earlier that the strata forming a fold always undulate along the strike. This is known as *undulation* of the axis inasmuch as the axes rise and fall together with the folds. Furthermore, no fold is infinite. An anticline usually terminates with a plunge of the axis while at the termination of a syncline the axis rises. In other instances, however, a fold terminates by the flattening out of the strata (that is the attitude of the strata becomes horizontal) or by attaching itself to another fold.

A fold is said to be *linear* if its length is many times its width. Short folds which are just a little longer than they are wide are

called brachy-folds, which in turn are divided into *brachyanticlines* and *brachysynclines.* As different from linear folds in which the axis over a considerable distance remains nearly horizontal, in brachy-

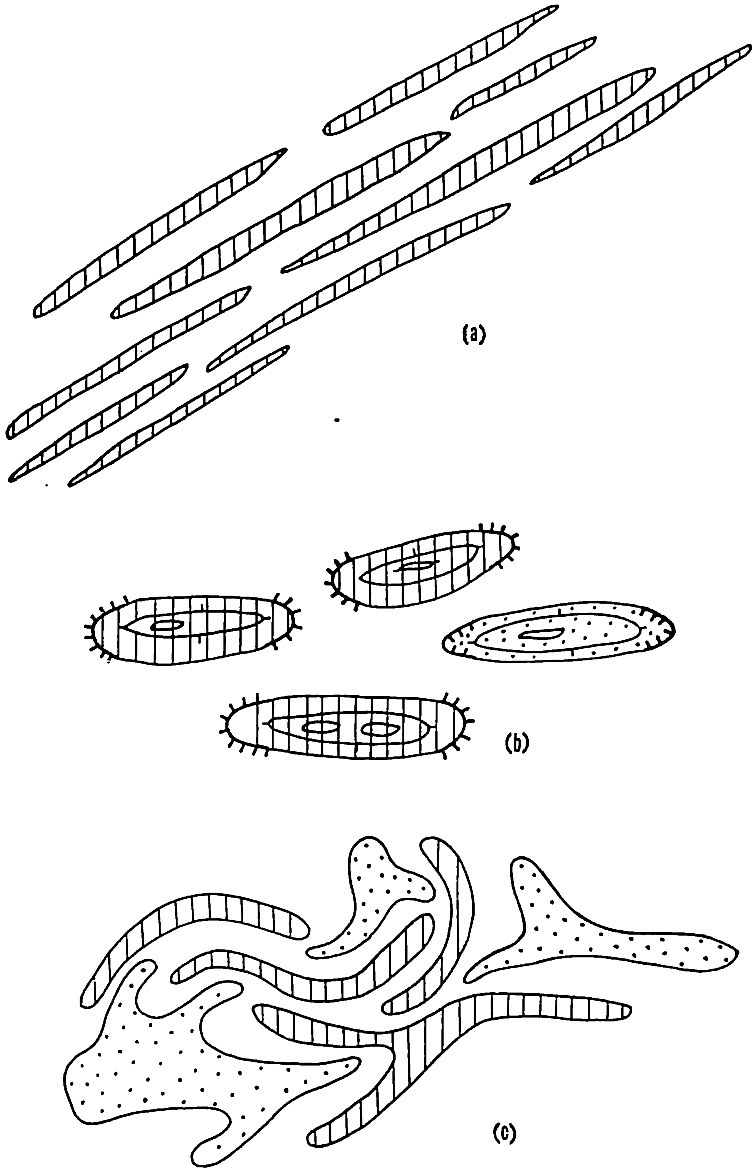


Fig. 45. (See legend on the next page)

folds it undulates much more strongly even on short stretches. A brachyanticlinal fold plunges in either direction along its strike, while a brachysyncline rises on similar short stretches.

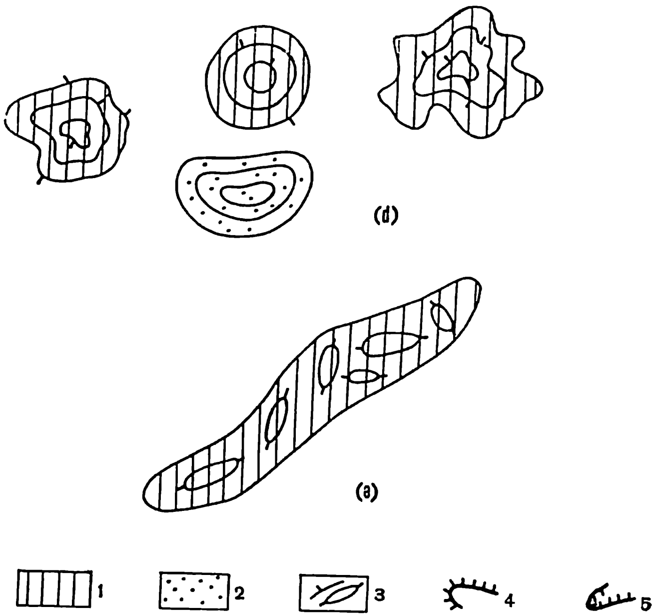


Fig. 45. Different fold shapes in plan:
a—linear folds; *b*—brachy-folds; *c*—arctuate folds and virgation (branching) of folds; *d*—domes and basins; *e*—elongated brachyline complicated by additional domes; *1*—relatively more ancient rocks in cores of anticlines; *2*—relatively younger rocks in synclines; *3*—symbolic stratoisohypses; *4*—periclines; *5*—centriclines

The strike of both linear and brachy-folds is not always straight—folds may be arcuate, broken and even sinuous.

Frequently folds branch out (this is also known as virgation) dividing along the strike into two or more diverging folds. Most often this occurs near the plunge of an anticlinal arch or the rise of the trough of a syncline.

The attitude of strata in the areas of plunge of linear anticlines and brachyanticlines and in the areas of rise of the axes of synclines and brachysynclines has certain specific features. Where an anticline or a brachyanticline plunges, the strata exposed at the surface describe an arc and radially dip away from the centre (Fig. 45*b*). Conversely, the strata of the trough of a syncline or brachysyncline also form an arc at the surface but dip radially towards its centre. Therefore, whereas along the limbs of a linear fold on a considerable

stretch the strata dip perpendicularly to the strike of the fold, at the plunge of anticlines and the rise of synclines the direction of their dip changes radially. This gradual change of dip links the opposite limbs of a fold. The mode of occurrence of strata at the plunge of an anticline (or brachyanticline) is called a *pericline* and that at the rise of a syncline (or brachysyncline) is called a *centricline*. Usually at periclines and centriclines the strata dip at a lower angle than at the limbs of the same fold. However, occasionally periclines and centriclines are steep and, in very rare instances, may even be overturned. In the latter case at the plunge of a brachyanticline the strata dip not away from the fold but beneath it. This may look as if it were not the plunge of an anticline but the rise of a syncline. Sometimes periclines and centriclines are referred to as closures.

Special types of folds are domes and basins. A dome is an anticlinal fold which is not elongated in any single direction in view of which one cannot speak of its strike. In plan such a fold appears as a rounded or irregular uplift. A basin is a similar, round or irregular, syncline (see Fig. 45*d*).

11. RELATIONSHIP BETWEEN FOLDS OF DIFFERENT STRATA

Up or down a vertical section of the earth's crust one notices that successive strata are not always crumpled in the same way.

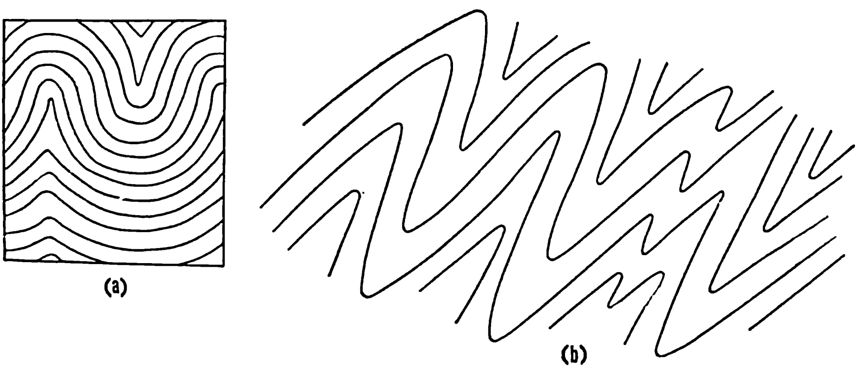


Fig. 46.
a—concentric folds; *b*—similar folds

If in a crumpled series the surfaces of strata form arcs that seem to have been drawn from a single centre, such folds are called *concentric* (Fig. 46*a*). Concentric folds most rapidly flatten out up or down the section and, therefore, never involve series of great thickness. Such folds are not common in nature.

More widespread are folds referred to as *similar* (Fig. 46*b*). In ideal similar folding the radius of curvature of the strata remains the same.

It will be seen from the graphic representation that this type of folding is possible only if the strata at the crests of folds are thicker than at the limbs. Furthermore, this variation in thicknesses depends on how steeply the strata dip on the limbs (the steeper the dip the

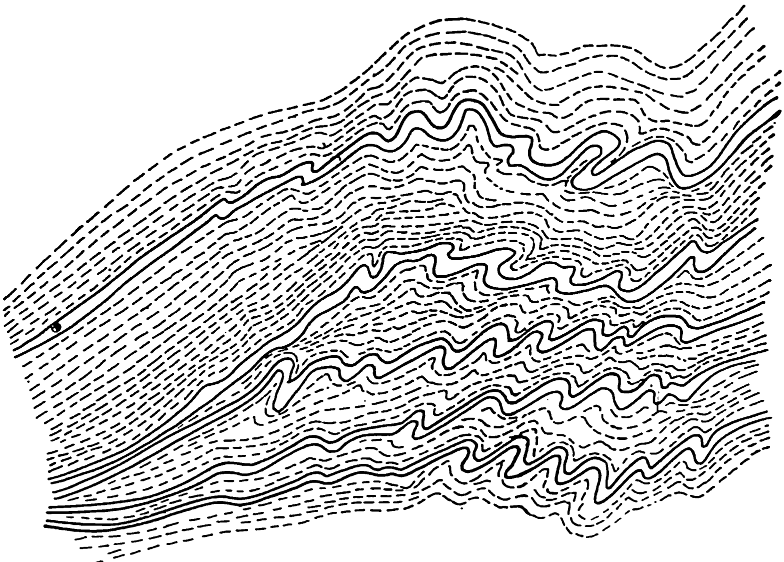


Fig. 47. Disharmonic folds in Archean rocks of Karelia (after A. A. Sorsky)

greater the difference in thickness between the limbs and the arches). Considering that originally over a short stretch corresponding to the width of several folds the strata probably had the same thickness, it must be assumed that the change in the thickness of strata occurred in the process of folding. This assumption is correct and we shall return to it later.

In similar and concentric folds considered above, strata deposited at different times were crumpled more or less conformably. In any case, anticlines in the upper layers have corresponding anticlines in the lower layers (the same applies to synclines). Such folds are usually called *harmonic*. Very often, however, such coincidence of attitude and similarity of folds in strata of different age is not observed. For example, often within the same series thicker and more massive strata are bent into broad gentle folds, while thin plastic layers are intensely folded. Such folds are called disharmonic and

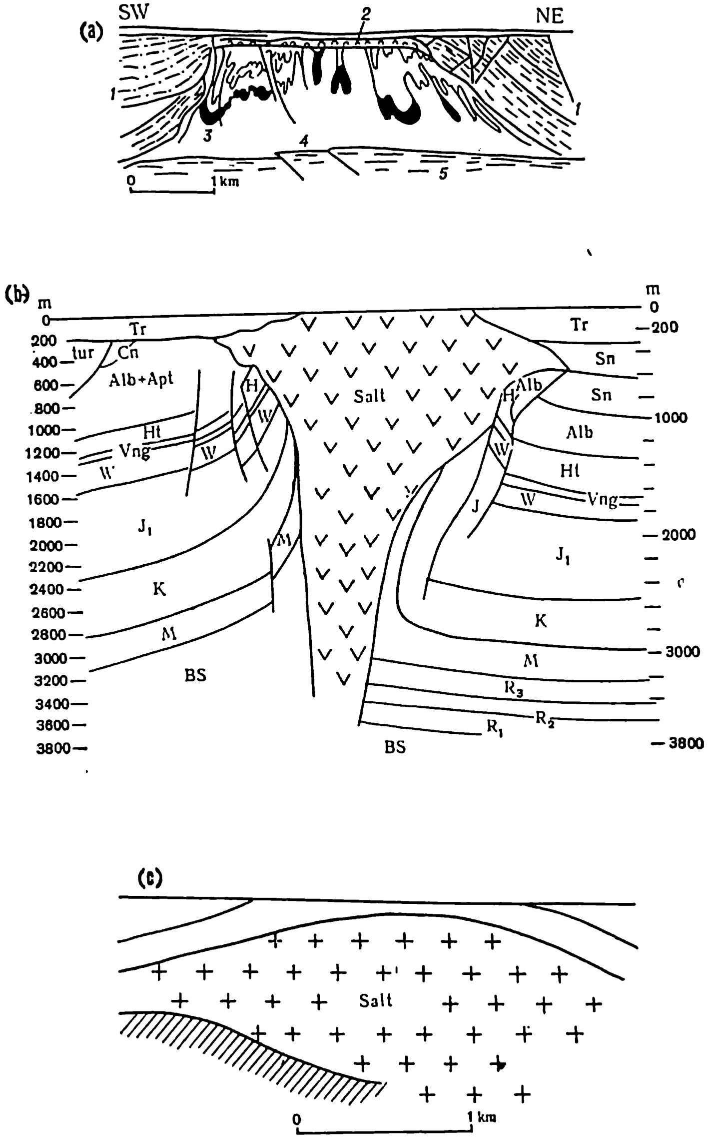


Fig. 48. Diapir domes with cores of different shape and composition:
a—cross section of a diapir dome in the FRG: *1*—variegated sandstone (Lower Triassic); *2*—caprock; *3* and *4*—anhydrite and salt (Upper Permian); *5*—more ancient Paleozoic basement; *b*—diapir dome with mushroom-like core (FRG); from BS to K—Triassic; J_1 and W—Jurassic; from Vng to Sn—Cretaceous; Tr—Tertiary; c—fold with lenticular diapir core (Ural-Emba region)

the differences in shape and other features (size, degree of compression, etc.) recorded in a single section are known as *disharmony* (Fig. 47). However, only those folds are termed disharmonic that were formed simultaneously, considering that the differences in the intensity and shape of folds may be due to the fact that folding in different series occurred at different times (of which more below).

It is evident that the development of disharmonic folds is associated with secondary changes in the thickness of strata. In some places they become thicker, and in others thinner, as a result of redistribution of the material of the strata.

Diapir Folds

Disharmony is most strongly expressed in the so-called diapir folds, also referred to as *diapir domes*. In these folds at least three different rock complexes characterized by different modes of occurrence are superposed on one another. The most important is the central complex composed of a highly plastic rock—salt, gypsum, clay, finely-laminated marls, or coal (Figs. 48, 49). In the process of folding this complex, which originally was of a uniform thickness on a considerable stretch, is squeezed out from some places which causes a corresponding considerable increase in thickness in other places. The swellings of this plastic complex are called *diapir cores*; they appear as lenses, more or less steep ridges, cones, cylinders or columns. Such columns sometimes are several kilometres high though only a few hundreds of metres in diameter. Some diapir cores form projections (ledges) intruding into the enclosing rocks, giving the core a mushroom-like shape. Diapir cores are most sharply expressed in those cases when the plastic complex is composed of salt. If the rocks of the plastic complex are laminated, the strata within the diapir core are crumpled into narrow compressed isoclinal folds.

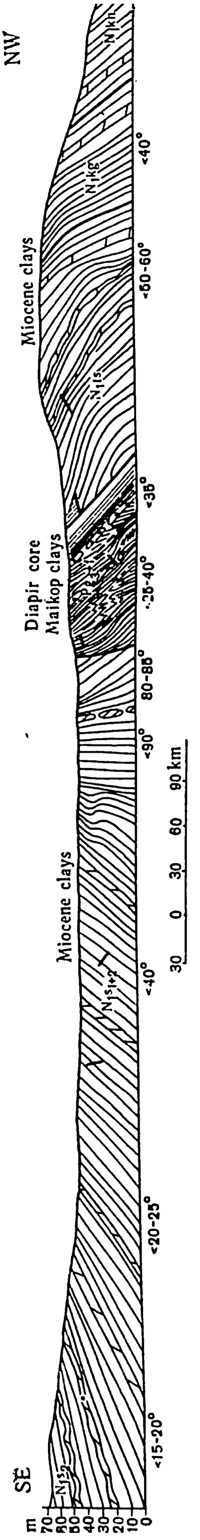


Fig. 49. Diapir dome with a clay core (Kerch Peninsula)

The rocks overlying the plastic complex are uparched above the diapir core and usually appear as a dome, complicated by ruptures. Often the rocks of the plastic complex intrude into the layers of the overlying complex, piercing and pushing them aside.

Because of that the diapir core is also referred to as the *piercement core.*

Generally little is known of the structure of the rock complex underlying the plastic complex. Available data indicate, however, that its mode of occurrence does not correspond to that of the plastic and of the overlying complexes. Sometimes the attitude of the lower complex is undisturbed, that is horizontal or gently monoclinic; in other cases it forms gentle uplifts and troughs of the dome and basic type, whose mode of occurrence does not correspond to that of the overlying complexes.

12. MORPHOLOGICAL TYPES OF FOLDING

Heretofore we spoke about the principal morphological kinds of separate folds. In nature folds do not occur singly but as large or small groups which occupy a certain area of the surface of the earth. The entire complex of folds found in a specific area constitutes what is known as *folding* or as a *folded structure.*

It is known from observations that the morphological features of folding may differ depending on the shape of the folds and their relationship both horizontally and vertically.

Depending on the shape and the areal relationship between folds two principal types of folding are recognized: *continuous* or *holomorphic* folding and *discontinuous* or *idiomorphic* folding.

Continuous or holomorphic folding is primarily characterized by continuous areal distribution of folds. In this folded structure the strata everywhere are crumpled into folds without any undisturbed areas in between. It may be said that this folding completely occupies the space within a certain region, which is then known as a *folded region* or *zone.*

Another feature of continuous folding is equal development of anticlinal and synclinal folds which are approximately of the same width and throw and also closely resemble each other in shape. This will become evident if the profile of a zone of continuous folding is reversed, so that the upper strata are below and the lower strata on top. In the case of typical continuous folding the general appearance of the zone will not change though anticlines will become synclines and vice versa.

Furthermore, in continuous folding the folds are linear and parallel. All the folds are extended and their length is many times their breadth. The strike of a considerable bundle of folds is the same and

within such a bundle the folds are parallel to each other. If the strike changes, this change simultaneously involves many or even a whole bundle of folds.

The last characteristic feature of complete folding is the uniform incline of the folds. Wide bundles of folds are inclined in the same

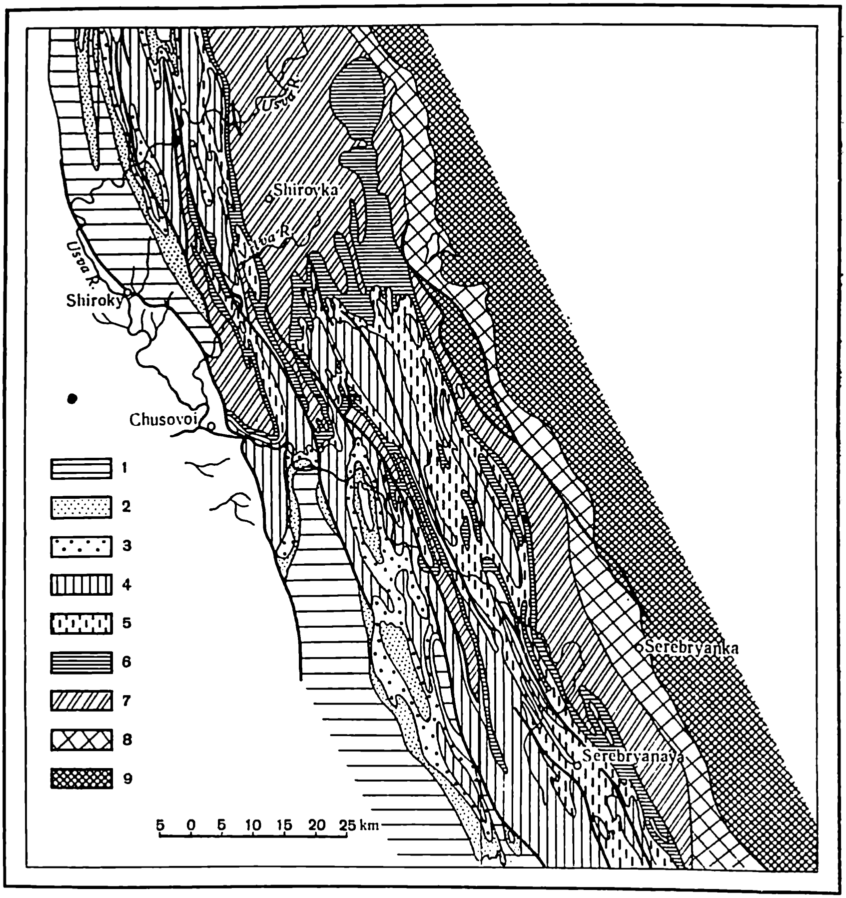


Fig. 50. Complete or holomorphic folding. Geologic map of a part of the Urals: *1*—Lower Permian; *2*—Upper Carboniferous; *3*—Middle Carboniferous; *4*—Lower Carboniferous; *5*—Upper Devonian; *6*—Middle Devonian; *7*—Lower Devonian; *8*—Upper Silurian; *9*—Cambro-Silurian

direction. Traversing a folded region at right angles to the strike of the folds, one may observe variation of the incline of the axial planes of the folds, but these just as the variations in the strike, always involve a considerable number of adjacent folds.

Examples of continuous folding are given in Figs. 50 and 51, representing a map and a profile of folded zones.

Considering the specific features of continuous folding, it may be said that individual folds do not constitute separate structural formations, but form parts of a folded structure and with regard to strike, inclination, and shape conform to the general pattern of folding characteristic of a given zone. Considering the continuity of the folds mentioned earlier, the impression is that a continuous folded structure develops as a result of a general crumpling process which simultaneously involves all the strata over a more or less considerable area.

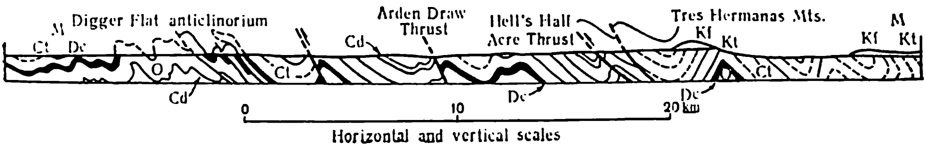
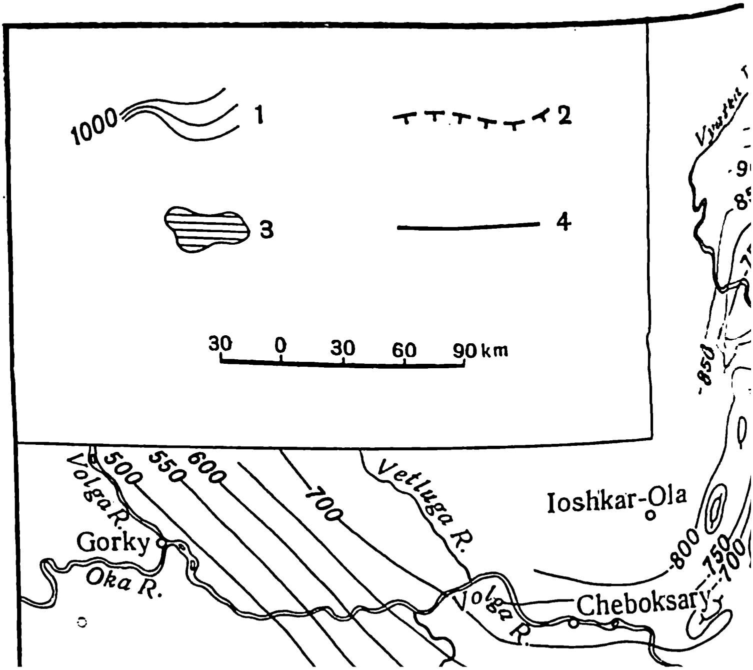


Fig. 51. Complete or holomorphic folding. Geological cross section of Marathon uplift (USA). Rocks from Cambrian to Carboniferous are involved in the folding. Cretaceous rocks occur unconformably and horizontally

Discontinuous or idiomorphic folding is characterized by features which constitute the opposite of those of continuous folding, primarily by the intermittency or the local nature of the folds. Most typical of discontinuous folding are isolated folds in a region generally characterized by horizontal position of strata. Another specific feature is the differential development of anticlines and synclines. In typical cases, anticlines are the only active forms without any corresponding synclines. In this case synclines are replaced by areas in which the strata have retained their horizontal position. The shape of such depressions is entirely controlled by the shape of the anticlines and their relative position. Where two isolated anticlines are very near to each other, a syncline is formed. Along the strike, however, such a syncline passes into horizontal strata as soon as the distance between adjacent anticlines increases or where they terminate.

Discontinuous folding is typically characterized by the absence of linearity. This is recorded in the fact that the isolated uplifts that form this folding are represented mainly not by elongated folds but by brachyanticlines and domes. If, however, long uplifts do occur among these forms their strike is not uniform and varies from one uplift to another. Only occasionally a large number of folds of the brachyanticlinal type exhibit a generally similar strike over a more or less considerable area. Even then, however, the independent nature of individual uplifts is underscored by the considerable



1000
1
2
3
4
30 0 30 60 90 km
Vyatka R.
-850
Volga R.
500
550
600
700
Vetluga R.
Ioshkar-Ola
Gorky
Oka R.
800
750
700
Volga R.
Cheboksary

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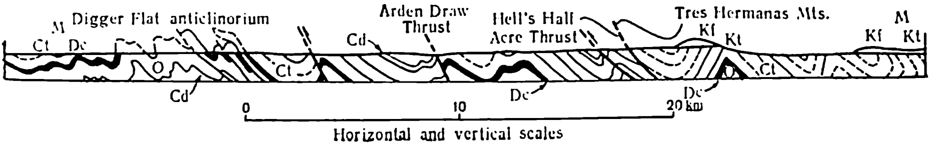

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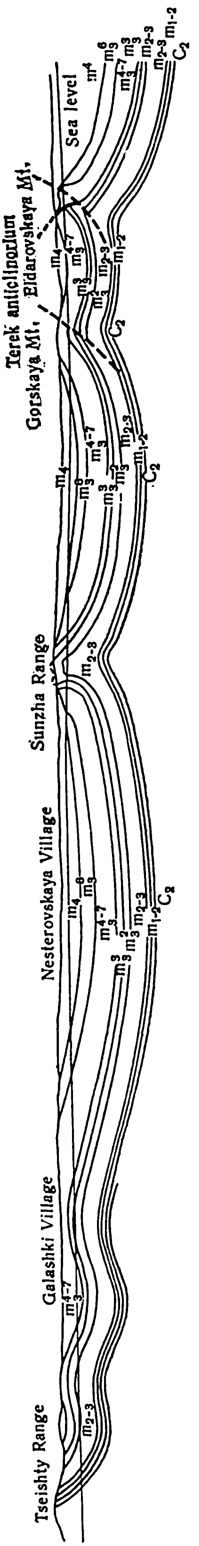


Fig. 54. Ridge-like folding; cross section of Tersk depression in Northern Daghestan (after I. O. Brod): m_4—Postpliocene and Pliocene; m_3^8—Chokrak; m_3^{4-7}—Sarmat; m_3^3—Karaganian strata; m_3^2—Chokrak; m_{2-3}—Maikop; m_{1-2}—foraminiferal strata; C_2—Upper Cretaceous

type. They are tens of kilometres in length and have a throw of hundreds of metres, sometimes more than a kilometre. The crest of the uplift is broad and flat and complicated by gentle domes of the second order. In contrast to the flat top, the limbs are steep and flexure-like. Sometimes one limb forms a flexure, while the other limb dips gently (unilateral coffer folds). In the typical case, the limbs are complicated by stepped flexures and, if the limbs are steep enough, by ruptures.

Discontinuous folds are also represented by elongated low uplifts. Their length is much greater than their width and the dip of the strata is usually negligible (a few degrees or a fraction of a degree). An example is the Oka-Tsna uplift which extends meridionally for 350 km from the Klyazma R. in the north to the Tsna R. in the south and is 25 to 40 km wide. The strata are thrust up for 200 to 300 m. The dip of the strata at the limbs does not exceed 1.5°. In plan the uplifts show a complex configuration—festoons, branches, "noses", etc.—and are always complicated by domes of the second and third orders.

Diapir folds also belong to the type of discontinuous or idiomorphic folding.

In addition to these two principal morphological types of folding, an intermediate type may be recognized. This, in turn, may be subdivided into ridge-like and coffer or box-type.

In typical ridge-like folding, strongly compressed anticlines (straight, inclined or fan-like) alternate with broad, flat synclines. The folding is thus characterized by unequal development of anticlines and synclines, but the synclines are not altogether missing as in the case of discontinuous folding (Fig. 54).

Box-type folding, as the term implies, is represented by numerous box folds. In

this case anticlines are the dominant feature. The shape of synclines is controlled by the shape and position of the anticlines. If these are near each other, the synclines between them are compressed and narrow, sometimes slit-like. If the anticlines are wide apart, the synclines have a flat floor and a general box-like profile.

In plan ridge-like and box-like anticlines appear as short oval brachyanticlines.

13. THE METHODS OF STUDY OF FOLDING

Folding is studied in the course of geologic mapping and are recorded on geologic maps and profiles of different scales. It is important, however, that all the principal features of fold shape should be recorded as accurately as possible even in general geological investigations. Often the shape of folds is sketched on profiles so schematically that the differences between continuous, ridge-like and box folds (i.e., those features the knowledge of which is absolutely essential to the understanding of the folding mechanism) are lost altogether. This is due not so much to the geologist's inattention to detail as to psychological factors. Accustomed to think in terms of the earth contraction hypothesis, geologists had long conceived folding as a regular succession of waves of strata. Inasmuch as ridge- or box-type folds do not fit into the contraction mechanism of folding, which allows for longitudinal bending only*, they were regarded not as regular phenomena but as freaks.

Studies of the mechanism responsible for a particular type of folding require special observations, more detailed than in general geological surveys. The object of these observations is to record on profiles, diagrams, sketches and other graphic material all the features of the folded structure, big and small, in their relationship, maintaining the same scale throughout. The last point is particularly important, considering that in ordinary geological investigations major dislocations are recorded correctly but very schematically, without observing the proper scale in sketching minor dislocations (usually it is exaggerated). This gives a distorted picture of the relationship between structural forms of different orders and of the conditions under which the dislocations in question had formed. Of course, it is not easy to plot on the same scale major folds and the fine folding that complicates them as well as boudinage and jointing. This must be done, nevertheless, adopting, if need be, different scales for the entire fold and for separate parts of it.

Only having carefully studied all the structural features of folds can the geologist draw conclusions on the mechanism and conditions

* Longitudinal bending develops under compression acting in the plane of a stratum.

of their development. Apart from establishing the time relation between the dislocations (for example, angular non-conformities, see Chapter VIII) it is very important to compare the character of deformations in different folds and in different parts of the same fold. Considering that dislocations do not develop uniformly, it is always possible to find and compare dislocations that are at different development stages. This can be done even more positively with regard to different parts of the same fold. For example, the central, most uplifted and the most strongly compressed part of a complete fold represents a more advanced stage of deformation than its periclines, whose development was arrested at earlier stages. By such comparison it is possible to reconstruct the picture of the movement of matter in a developing fold.

In recent years, the method of investigating the mechanism of folding on models is being extensively used in the USSR. It consists of reproducing in the laboratory on a reduced scale folds similar to those observed in nature. In this way it is possible to observe many new dependences and features of the folding process which elude the geologist in the field because in nature the geologist does not see the deformation process and necessarily must confine himself to studying its end result.

Tectonic modelling is based on the principle of physical similarity. Two physical processes are similar if they are identically described by similar dimensionless equations. This is possible when a definite relationship is maintained between the values of physical parameters that enter into these equations. Such relationships are referred to as the conditions or criteria of similarity. It has been established that for modelling slow plastic deformations (such as folding) when the inertial forces and elastic phenomena may be disregarded, the conditions of similarity will look as follows:

$$C_\eta = C_\rho C_e C_t,$$

where C_η = ratio between the viscosity values of the model and the object being modelled,

C_ρ = ratio of densities of the model and the object,

C_e = ratio between the geometric dimensions of the model and the object, and

C_t = ratio between the time of deformation in the model and in nature.

It will be seen from the equation, that three out of four physical parameters of the model may be chosen arbitrarily, but this choice will determine the fourth, which cannot be arbitrary. It also follows from the same equation that if in modelling we scale down considerably the dimensions of the object and the time of deformation, the viscosity of the modelling materials should be less than

that of the object. In view of that rocks are modelled by various soft, even flowing materials (wax, soft clay, etc.).

Technical factors (size of model, time required, the possibility of making sections and recording the model structure, etc.) to a certain degree restrict the choice of modelling materials. Calculations show that the most suitable materials for modelling slow plastic deformations in thick series of sedimentary rocks, for models 5 to 20 cm thick in experiments lasting less than 24 hours, are certain petroleum products (Baku petrolatum, gun grease, bitumen, rosin or moist clay).

For ideal modelling it is essential to know the viscosity of rocks under the conditions of their deformation in the earth's crust and the duration of natural tectonic processes. Actually, however, these are known only in the most approximate way. It is true that viscosity can be investigated in special laboratories, but it can hardly be expected that our knowledge of the rate of tectonic processes would become much more accurate in the near future. Of course, this detracts from the accuracy of tectonic modelling. Nevertheless, the application of modelling to the study of various tectonic dislocations shows clearly that despite this inherent inaccuracy tectonic modelling may be regarded as a valuable investigation method. Though not independent, it is most important as a complement to geological field investigation methods. When correctly applied it permits to record certain important features of the process being reconstructed.

The use of models is a most important part of the complex of the so-called tectonophysical investigations. Tectono-physics is the science that studies the physical mechanism of tectonic dislocations, using both geological and physical methods. To obtain a correct picture of the development of a particular dislocation it is necessary to reconstruct, as far as possible, the stress field that existed at that time in the earth's crust due to the action of tectonic factors. This can be achieved on the basis of study of fissuring and faulting (see Chapter VI). Various tectonites may be used for the same purpose (see the same chapter). But tectonic stress fields can also be reproduced on models by the so-called photoelasticity method using special optically-active transparent materials (gelatins). Subjected to deformations similar to those observed in nature, a model made of such material develops stresses. The model is illuminated with polarized light on an installation similar to a petrographic microscope. The distribution of shear stresses of different intensities is established either by means of an optical compensator or directly from the interferential colouring of the image of the model projected on a screen.

CHAPTER IV

The Folding Mechanism and the Internal Structure of Folds

14. KINEMATIC CLASSIFICATION OF FOLDS

In the preceding chapters we have described the principal morphological types of folding. On the basis of available data it is also possible to discriminate different types of folds by the mechanism of their development. The folding mechanism is the movement of the earth's crust or of the rocks composing it that directly produced the given fold.

Strata may be bent into folds either by the so-called transverse bending or by longitudinal bending. In the first case the stratum is bent under the action of forces that are applied to it normally, that is in a direction perpendicular to its plane. Inasmuch as originally the attitude of strata is mostly horizontal, the forces that cause transverse bending are directed vertically. Where they act from the bottom upwards the strata are uplifted and bent into anticlinal folds.

In the case of longitudinal bending the stratum is subjected to the action of horizontal forces directed along its plane. In this case the stratum is bent into folds by compression. The shape and dimensions of anticlines and synclines and their distribution is then controlled chiefly by the reaction of the stratum to mechanical stress, that is by the mechanical properties of rocks.

The most common examples of transverse folds in nature are discontinuous and box-like folds, though in some instances they may have been formed by a different mechanism.

In typical cases, discontinuous and box-like folding develops under the action of upward tectonic forces and is expressed locally in separate districts. The folding thus produced may be termed *block folding*, since in such cases the bending of strata in the upper parts of the earth's crust is often due to vertical displacement of deep-seated blocks bounded by faults (Fig. 55*a*).

Folding due to longitudinal bending is usually expressed in more or less parallel crumpling of a very thick series of strata. In view of that, folding of this type may be termed as *general crumpling* (Fig. 55*c*).

These two types are distinguished by the differences in the folding mechanisms, and therefore form elements of a kinematic classification of folds.

A special type f injection folding is recognized as well. It holds an intermediate position between the two other types of folding

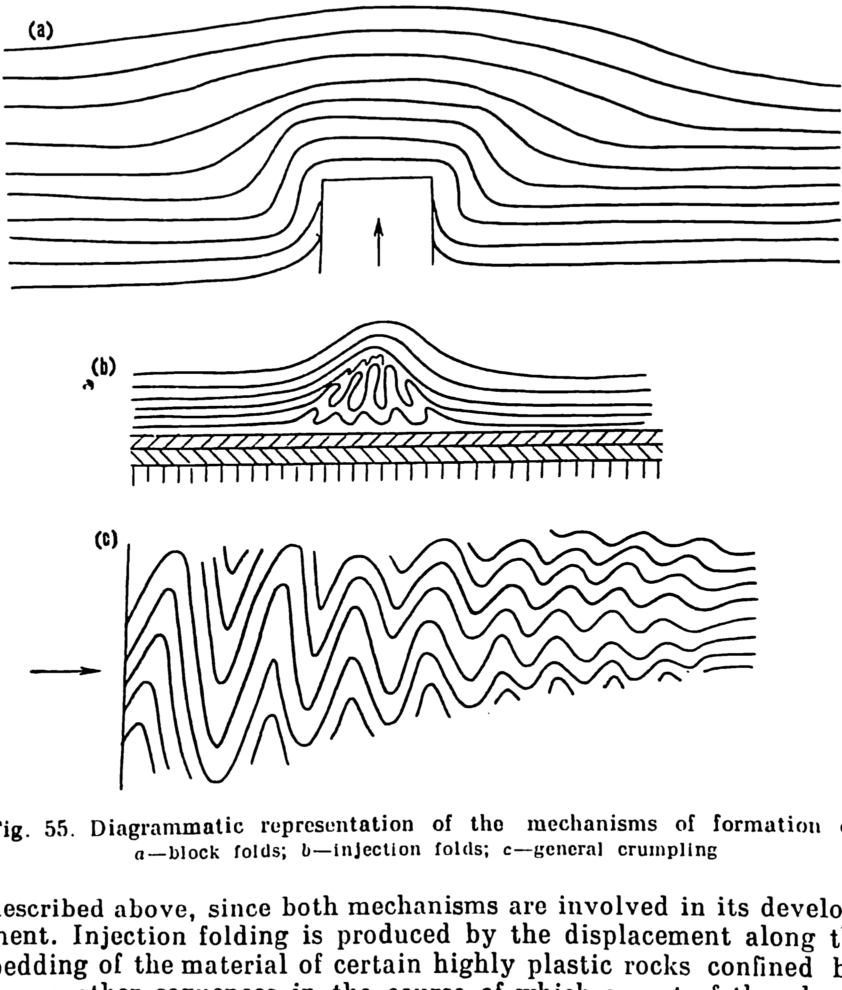


Fig. 55. Diagrammatic representation of the mechanisms of formation of: a—block folds; b—injection folds; c—general crumpling

described above, since both mechanisms are involved in its development. Injection folding is produced by the displacement along the bedding of the material of certain highly plastic rocks confined between other sequences in the course of which a part of the plastic material is squeezed from some places and injected into others. Where the plastic material accumulates the sequence becomes thicker and the overlying rocks are uparched, forming, in the general case, an anticline (Fig. 55b). Thus, this type of folding is characterized

by the division of the rock series into an active or mobile and passive complexes. The active complex is confined between the passive and the movement that develops within it affects the attitude of the overlying rocks. Thus, longitudinal compressive forces are active within the mobile complex in the regions of injection of material, as a result of which in these regions the strata of the active series are often extended and compressed into folds.

The active series of very plastic rocks most often is composed of salt, gypsum or soft clay. On a small scale injection folds are also formed in thick seams of coal, which under tectonic conditions become highly flowable.

The upper complex is bent by upward pressure exerted by the material of the active complex. Thus, the upper complex is subjected to the action of the mechanism of transverse bending. The rocks underlying the active complex do not suffer deformation.

To this kinematic type of folding belong the diapir folds described above and apparently all, or most, of ridge-like folds.

Now we shall consider certain details of folded structures of different kinematic types.

15. BLOCK FOLDING

Morphologically block folding is expressed, in particular, as discontinuous folding, whose principal features were examined earlier.

Certain interesting structural features of discontinuous folding are revealed when its section is investigated in depth. Most often the series involved in folding of this type exhibits regular thinning of the strata towards anticlinal crests and thickening of the strata in the depressions between the uplifts. As a result the deeper layers are most sharply bent than those near the surface. These variations in thickness are associated with the development of discontinuous folds. Inasmuch as the thickness of sedimentary series is controlled by the magnitude of the subsidence of the earth's crust, the thinning of strata at the crests of folds indicates that these areas subsided more slowly than the adjacent areas (now occupied by the depressions between the uplifts), and that discontinuous folds developed as a result of this non-uniform subsidence.

Certain horizons or even whole systems, not only diminish in thickness but even completely pinch out at anticlines. The rate of their deposition in synclines corresponds to the rate of erosion on the crests of anticlines, which, therefore, experience absolute uplifting at such periods. Thus, the history of discontinuous folds consists of alternating periods of relative and absolute uplifting of anticlines.

In some cases, however, this picture of variation of thicknesses within individual folds is not so simple. Sometimes, for example, starting from a certain stratigraphic level the thickness does not change with depth. Conversely, in other cases, starting at a certain depth, the thickness of the series increases towards the crest of an discontinuous anticline as a result of which the anticline flattens out with depth. In places the anticlinal attitude of the strata may pass into horizontal or monoclinal; at still greater depth it may become synclinal. Lastly, in some cases the distribution of thicknesses is irregular which results in considerable discordance of structural forms at different depths.

Considering what has been said above concerning the conditions that control the thickness variations, it may be concluded that all the cases mentioned above attest the complex history of crustal movements. If thickness increases towards the crest of a contemporary anticline, it is to be inferred that at the time of deposition of the series in which such an increase is observed the given section of the earth's crust was involved in subsidence. Later on it must have been replaced by uplifting during which the distribution of thicknesses conformed to the usual pattern, that is decreased towards the crest of the anticline.

In some cases the distribution of thicknesses in areas of uplifting and subsidence might have changed repeatedly. These changes are recorded mainly in the fact that at different horizons reduced or increased thicknesses are observed at different places.

Thus, the present distribution of discontinuous folds and their shape is the result of a very complex succession of up and down crustal movements.

When a discontinuous anticline develops in a series of strata whose thickness varies in a regular way areally (increasing in a direction transverse to the strike of the fold) the position of the anticlinal crest at different levels of the vertical sections is displaced. In higher horizons the crest is displaced in the direction of increasing thickness.

The discordance in the bedding of different horizons involved in discontinuous folds has very important implications. It should be reckoned with, for instance, in drilling for oil and gas which tend to accumulate in the crests of anticlines, whose position at depth may be not what it is at the surface. Oil and gas accumulations may be entirely absent at depth in those areas where an anticline flattens out or is replaced by a syncline.

Modelling of block folds has shown that their flattening out from the bottom upwards, is associated not only with non-uniform accumulation of strata but also with a purely mechanical process determined by the viscosity of the material. Pronounced folds of the

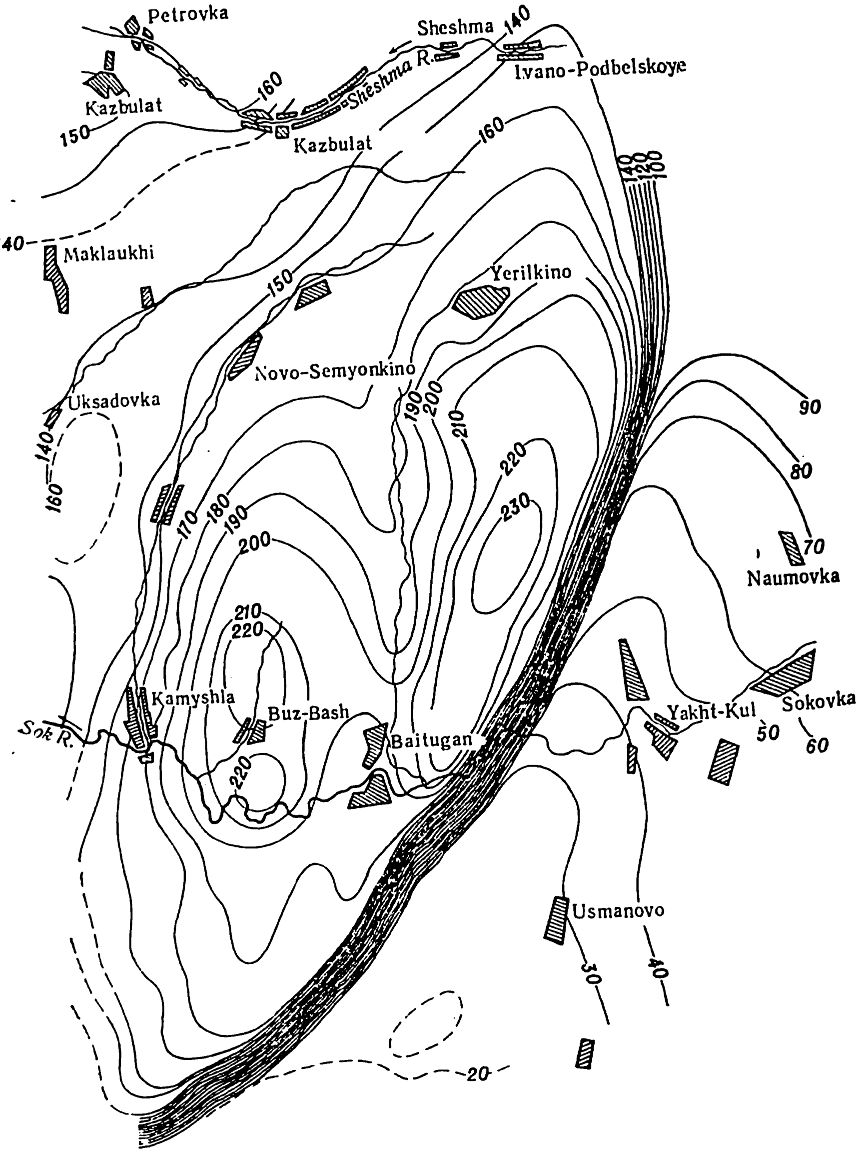


Fig. 56. The Baitugan rampart (Tatar Autonomous Republic) complicated by small domes (stratoisohypses are plotted along the roof of the spiriferal horizon)

box type observed in depth (anticlinal and synclinal), in the upward direction pass into very gentle and wide upwarps and downwarps. It is even possible that two separate uplifts observed at depth may merge into a single gentle brachyanticline at the surface.

These observations confirm that it is possible to regard discontinuous folds as block folds: the insignificant upwarping of strata at the surface in the form of a brachyanticline or dome may be the expression of the upthrusting of a deep-seated block bounded by faults. In view of that brachyanticlines and domes of the discontinuous type are sometimes referred to as "reflected block folds" to

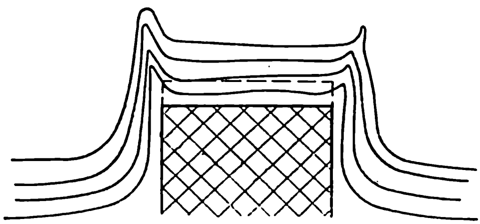


Fig. 57. Additional folds over the limbs of a box fold. Dashed lines designate the earlier higher position of the block whose rise caused the development of the fold

distinguish them from block folds proper, whose structure more clearly shows their connection with the upthrusting blocks of the earth's crust.

In the typical case major discontinuous folds are complicated by smaller discontinuous folds or domes (Fig. 56).

An interesting point is that box folds are occasionally complicated by longitudinal folds of the second order, which project above the flexures bounding the fold. Presumably, such projections could have developed as a result of subsidence of the block forming the core of the fold, after its initial upthrust. When the block was upthrusted the overlying strata suffered plastic stretching. When it slightly subsided, the surplus was bent into acute folds above the flexures (Fig. 57). A more plausible explanation is that these projections represent additional injection folds, produced by upward squeezing of material from the strata of the steep limbs of the flexures.

Occasionally discontinuous folds are complicated by smaller folds which may be classed as continuous or ridge-like. The throw of such folds does not exceed a few tens of metres or even a few metres. In such cases isolated bundles of small linear parallel folds fringe the discontinuous uplifts, tending to concentrate at their limbs, generally parallel to the strike of the strata (Fig. 58).

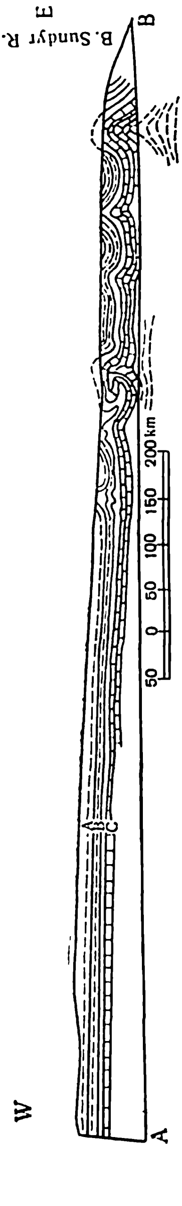


Fig. 58. Small secondary folds at a limb of a gentle dome in the area of B. Sundyr R. (after V. Bronguleyev). Letters denote various horizons of the Tatarian suite

Presumably, such small linear folds which complicate the limbs of discontinuous uplifts have been formed by the action of the injection mechanism: when discontinuous folds (domes) are uplifted, the strata at the crest of the fold, meeting resistance from the overlying layers, are compressed in the vertical direction. This compression causes flowage of the material of the most plastic layers from the crests and its injection into the limbs, where it accumulates and is crumpled into longitudinally-bent linear folds. However, such small folds may have also developed as a result of flowage of the plastic layers under the action of gravity or due to the squeezing out of this plastic material in gulleys, valleys and other depressions under the weight of the overlying rocks in adjacent higher areas.

In the course of uplifting of discontinuous folds of the box type, with steep flexure-type limbs, the strata are subjected to transverse compression and longitudinal extension not only at the crest but also on the limbs to the point that some layers may completely disappear.

The elevation and subsidence of separate blocks of the earth's crust is caused by internal subcrustal processes whose nature is still debatable.

16. GENERAL CRUMPLING

Three mechanical elements are involved in the development of general crumpling. In some cases a definite succession exists between these elements but occasionally they occur simultaneously. The first such element is the bending of strata involved in the folding. This bending is possible only because every stratum possesses a certain degree of freedom in the process of deformation, permitting it to slip relatively to the adjacent strata.

Let us imagine an isolated rock stratum in the form of a thin slab. Such an isolated layer relatively easily crumples into folds under

the action of horizontal compression. The thicker is the layer the more it resists crumpling by horizontal compression. Naturally a monolithic series of great thickness cannot be crumpled into folds at all. These conditions were simulated on a model (Fig. 59*a*) in

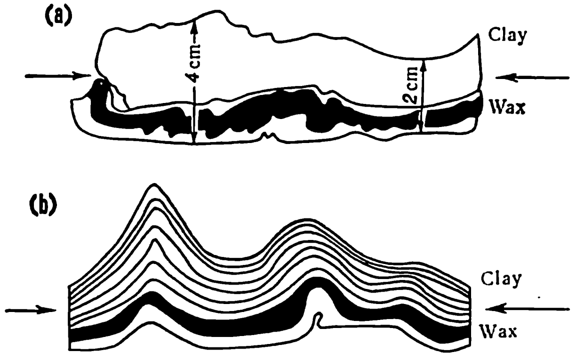


Fig. 59. The part played by layered structure in the development of folds:
a—deformation of monolithic clay layer; *b*—deformation of a layered clay packet

which a relatively thick monolithic layer of clay overlying a wax layer was subjected to horizontal compression. In this case horizontal compression did not produce regular folds, and deformation was expressed in a general increase in the thickness of the sample and its irregular warping.

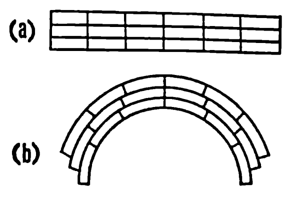


Fig. 60. Gliding of strata on bending:
a—initial position of strata; *b*—position of bent strata

Figure 59*b* shows the results of an experiment on another clay model of the same thickness. This time, however, it was separated into layers, the cohesion between which was weak. In this case compression did produce folds. These differences in the behaviour of the two models are explained by the fact that folds would not form in a series of layers unless the layers move on one another, which is known as interstratal gliding.

Indeed, let us visualize three superposed layers (Fig. 60*a*). The layers have marks which are in alignment. When the layers are bent into a fold only the marks in the arches of the layers will remain in alignment; on the limbs the marks in each overlying layer will be displaced relatively to the underlying layer (Fig. 60*b*). This dis-

placement attests to the gliding of layers along the bedding planes. Were the layers glued together it would be exceedingly difficult to bend such a series into a fold.

The thinner are the layers the more easily they are bent. Because of that in thin-bedded series (all other conditions being equal) the layers are compressed into more complex and sharp folds than in series composed of thicker layers. Another factor affecting the shape of folds is the strength of the rocks. In clay rocks folds are more sharply compressed and smaller than in hard sandstones of the same thickness.

In the case of sequences composed of layers of different thickness, the general shape of the folds is determined by the aggregate mechanical properties of the series.

If certain rocks predominate and the others play a subordinate role (for example, a clay series with thin sandstone intercalations) the former are the "dominant" layers that control the shape of the folds. The subordinate layers are forced to assume this shape even though it may not be typical of them. Often certain parts of a sedimentary series composed of layers of a different nature are differently crumpled which results in disharmonic folding.

Bending alone (of course with gliding of the strata) may result only in concentric folding. However, in nature much more common is the so-called *similar folding*, characterized by different thicknesses of the folds at the limbs and the crests due to redistribution of the material in the layers. The layers that are subjected to bending are crushed on the limbs and the material is forced (injected) upwards and downwards into the arches of the anticlines and synclines (Fig. 61). This occurs because on lateral (horizontal) compression of a stratified series by tectonic forces, as the strata are more strongly folded, the increasing component compresses the layer on the limbs not laterally but normally to the layer, squeezing its material upwards and downwards into the arches of the anticlines and synclines.

The extent of this squeezing varies with the plasticity of the rocks and, naturally, with the general intensity of compression. Highly plastic layers suffer greater deformation and may be completely squeezed out from the limbs, whereas the thickness of harder layers may change very little, or not at all. The American geologist Cloos tried to determine the magnitude of secondary variations of the thickness of strata on the limbs by changes in the form of such elements of rocks, whose original shape is known. As the most convenient objects for such observations he chose oolitic limestones, whose grains originally are more or less spherical. Studies of secondary changes of the shape of oolites in different zones of the Appalachian folding enabled him to distinguish zones marked by different degrees of the crushing of strata.

Only the most plastic strata thin out uniformly on the limbs of folds on crushing. More rigid strata suffer boudinage, that is, are separated into thick lenses with narrow necks in between. Furthermore, usually the lenses of one stratum adjoin the necks of the adjacent stratum and vice versa. This produces a peculiar wave-like lamination (Fig. 62). The size of the lenses depends on the thickness of the stratum. They are larger in thicker and stronger strata. The

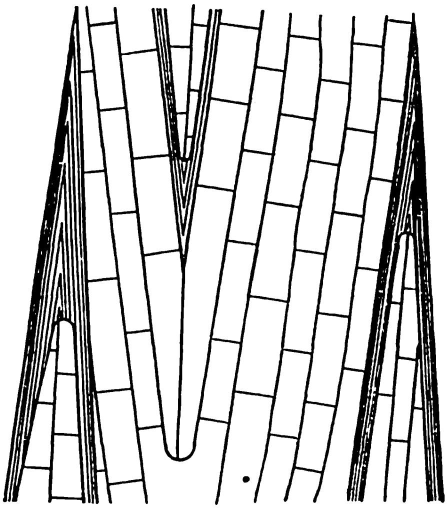


Fig. 61. Forcing of the material of crushed layers into crests of folds

size of the lenses also depends on the general deformation of the stratum in the given direction and varies (in length) from a few centimetres to one or more metres.

Boudinage results from the concentration of plastic gliding when, at regular intervals, a stratum is subjected to transverse crushing and longitudinal extension. Concentration of plastic gliding was discussed earlier in connection with the mechanism of viscous fracturing. The distributed nature of application of the crushing forces results in a similarly distributed extension of the stratum, which means that extension occurs at every point of the stratum. As a result it separates into a multitude of segments each with its own "neck". The size of such segments depends on the degree of deformation and the strength of the stratum, which in turn is determined by the mechanical properties of the rock and stratum thickness.

The lenses are ellipsoidal bodies whose axes lying in the plane of the stratum usually differ in length, being shorter in the direction

of the greatest deformation. Since the stratum is usually subjected to greater extension along the dip than along the strike, the lenses usually are elongated along the strike of the stratum.

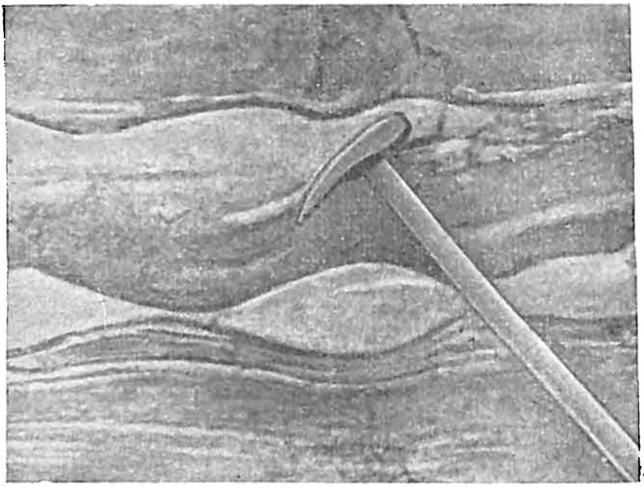


Fig. 62. Boudinage in Archean rocks of Karelia (after A. Sorsky)

If the series is not homogeneous and some strata are less plastic than others, their deformation lags behind the deformation of other

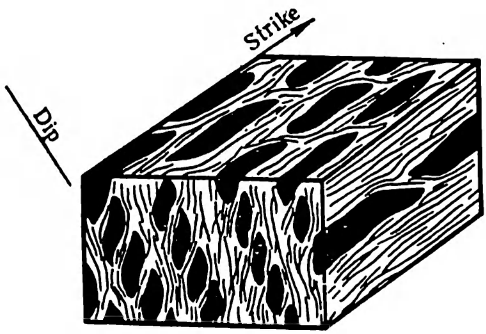


Fig. 63. Rupturing of limestone strata into blocks in Devonian marls of the Karatau ridge (sketch)

strata. The material of the more plastic strata is deformed at a quicker rate, spreads over the surface of the more rigid layers and stretches them. Under such conditions the lenses of the less plastic strata may become detached from each other along viscous shear fractures (Fig. 63). Considering that shear fractures are inclined

relatively to the stratum, and that the detached lenses are also slightly inclined, this may cause errors in estimating the attitude of the stratum. The intervals between the lenses are filled with more plastic material from the adjacent strata.

If a stratum of low plasticity or one that is almost incapable of suffering deformation under the given conditions occurs between plastic strata, by the same stretching process it is split into blocks which are dragged apart. Inasmuch as this separation takes place along tension ruptures, the blocks are parallelepiped in form, with sides perpendicular to the stratum. The dimensions of the blocks, just as the dimensions of the lenses, depend on the magnitude of deformation in the given direction, rock strength and the thickness of the stratum. The interstices between the blocks are filled with plastic rocks from adjacent strata. Mineral segregations (quartz, calcite, etc.) are often observed in such interstices as well. These segregations have formed in the zone of relieved pressure ("potential cavities") between the blocks that have moved apart. The rounded corners of the blocks indicate that before they separated small necks had been formed (Fig. 64).

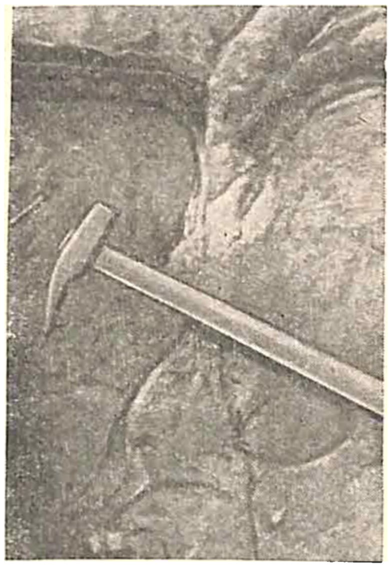


Fig. 64. A joint between two amphibolite blocks filled with pegmatitic material (after A. Sorsky)

Considering the more intensive deformation of plastic strata which is observed on the limbs and results in the rupturing of rigid strata, it is necessary to reconsider completely some earlier views on the mechanism of crumpling of strata into folds. At the end of the last century, the American geologist B. Willis advanced the theory of "competent" (more rigid or less plastic) and "incompetent" (more plastic) strata. He believed that tectonic pressure was transmitted through competent strata, whose bending determines the "skeleton" of the folded structure, while the bending of the incompetent strata was passive and subordinated to the bending of the more rigid rocks. He assumed that material was forced into the crests only in the incompetent strata as a result of their crushing between the competent strata.

However, facts disprove this view. In the development of complete folding it is the most plastic, the most flowable strata that are the most active. When such strata predominate in a structure, they control the form of the folds while more rigid strata, confined between them and separated into blocks, naturally could not serve as the "skeleton" of these folds.

Occasionally additional minor folds develop in the crests of the folds into which the material from the strata is forced and where therefore conditions exist for lateral compression relative to the layers. Such additional minor folds may also develop on the limbs of major folds as a result of local "overrunning" of material as it is squeezed out.

The next and last element of the folding mechanism is the general compression of the entire series involved in the crumpling.

With continued folding the folds become more and more steep and compressed. Under the action of the compressing forces the material of the strata gradually becomes indurated and friction at the bedding plane increases. Eventually further compression of the folds becomes impossible and, if horizontal compressive forces continue to operate, the entire crumpled series reacts to these forces as a solid, unstratified mass. At this stage stratification loses its mechanical meaning. Under these conditions the entire crumpled series is compressed as a whole in the horizontal direction, while its vertical thickness correspondingly increases. This stage is marked by the development of the so-called *flow or basal cleavage.*

At the surface of outcrops flow cleavage is expressed as very fine schistozity. The rocks are split into thin plates usually oriented parallel to the axial plane of the fold. Microscopic studies of cleavage show that it is associated with plane-parallel orientation of the minerals. Cleavage is most apparent in clay rocks, but is observed in other rocks as well. In alternating beds of different composition it is more pronounced in more plastic rocks, while in adjacent stronger rocks it may be absent or be replaced by jointing.

The plane-parallel orientation of minerals which determines the development of cleavage is associated with the crushing of minerals under the action of tectonic compression in a direction normal to the cleavage planes. As a result of crushing the minerals are flattened.

The fact, that flow cleavage is due to compression of a rock in a direction normal to the cleavage with corresponding elongation parallel to the cleavage, is confirmed by the character of deformation of various inclusions in rocks. Fossil inclusions in a cleavaged series are flattened along the axis normal to the cleavage and stretched along this axis. A classical example are the belemnites described by A. Heim in the 1870s. In the process of elongation these belemnites were torn into several pieces along the cleavage. Wetstein devoted

a large study to the deformed imprints of fish and confirmed that in strata ruptured by cleavage stretching occurs parallel to cleavage, and compression normal to it. The same dependence is evident wherever the position of deformed oolites can be related to flow cleavage (Fig. 65).

The fact that in flow cleavage the strata no longer slide relatively to each other is evident from the character of the bedding planes in

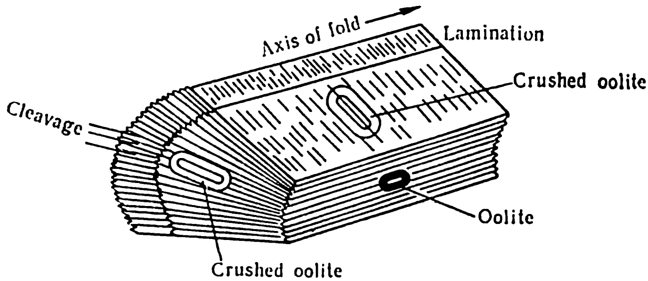


Fig. 65. Flattened oolites in folded and cleavaged rocks (after Cloos)

cleavaged series. When cleavage cuts across various strata the boundaries between them show a kind of serration. The separate plates produced by cleavage are displaced relative to one another so, that the serrations of one stratum engage those of another. Such "engagement" of the strata convincingly proves that sliding of strata no longer occurred in cleavage. The entire series of rocks in this case was deformed uniformly without sliding or bending of the strata and without differential flowage of the material of the strata. The serration of the bedding planes indicates that in the course of general compression the separate plates produced by cleavage suffered crushing to a different degree.

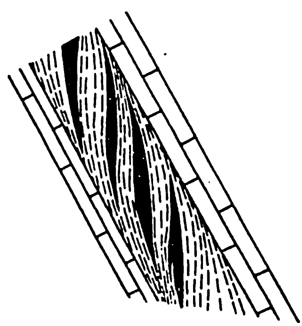


Fig. 66. Sigmoid cleavage

However, some observations indicate that, if conditions change after the development of basal cleavage, the relative movement of the strata may resume. In that case the cleavage planes are broken at the boundaries between the strata and are bent due to the friction of the strata against one another. It is in this manner that sigmoid cleavage is produced (Fig. 66).

The three elements of the folding mechanism in the typical case follow each other in the order described as stages of a single process. More often, however, they overlap and develop (at least in part)

simultaneously. This applies to bending and redistribution of the material in the strata which commonly occur together. Flow cleavage is observed only in regions of very intensive folding.

Apart from flow cleavage, so-called *fracture cleavage*, the origin of which is not certain, is also observed in folded series. Like flow cleavage, it is recorded in the division of rocks into plates which, however, are much thicker and coarser than in the case of flow cleavage. Fracture cleavage does not cut across series of strata but is expressed in each stratum separately. The frequency of cleavage planes as well as their inclination change from one stratum to another. Most often fracture cleavage is perpendicular to the stratum, but different orientations, such as a fan-like arrangement relatively to the axial plane of the fold, are known as well.

Possibly fracture cleavage is due to deformations developing in the strata in the course of their relative movement. When the strata slide (in connection with bending) each stratum sandwiched between the overlying and underlying strata is subjected to the action of a couple of forces and to shear deformation. Presumably fracture cleavage develops in conformity with the direction of maximum shear stresses. The fractures developing along one of the planes are parallel to the strata and merge with the bedding planes and are, therefore, unobservable. Fracture cleavage develops along another system of shear fractures, perpendicular to the plane of the stratum. This problem is not yet completely solved.

The origin of the horizontal forces that cause general crumpling is still far from being understood. Recently it was believed that this type of folding was due to the general contraction of the earth's surface (the contraction hypothesis). By now, however, this hypothesis has been rejected. According to present views the most probable cause of complete folding is the flow of stratified material of the earth's crust from uplifts into depressions with local damming on the way. The most likely cause of this flow is gravity, the crushing of relatively deeper strata at the crests of upwarps, and the resulting resistance of the overlying strata to this uparching.

17. INJECTION FOLDING

The movement of very plastic material within the earth's crust from a certain area towards one place, where an injection core or even a piercement core is formed, is due to various factors. One of them is the low specific gravity of a plastic series as compared with the overlying rocks. Such is the mechanism by which most salt diapir domes are formed. Rocks may be regarded as very viscous fluids capable of very slow flowage. A thick salt bed underlying rocks of greater specific gravity (the sp. gr. of salt is 2.13, the average sp.

gr. of fragmental rocks is 2.3) tends to float up and assume a position above the heavier rock. This occurs by intrusion of salt columns into the overlying rocks. Such injection folds may be called folds of *gravity buoyancy*.

Only few salt flows are observed on the surface in nature. One such case is known, for example, in Iran, where "salt gletchers" appear on the surface in connection with diapir domes. Commonly the movement of a salt stock stops at a certain depth below the surface.

The geological history of salt domes shows that the growth of each dome was repeatedly terminated to resume later. Possibly this is due to the fact that rocks possess not only viscosity but also strength which the rising salt stock must overcome, since its upthrusting causes the splitting of the overlying series. If the thickness of the overlying beds increases (as a result of new deposition) their total strength also increases, which may cause an interval in the rise of the diapir dome. When these deposits are partly eroded the growth of the dome may resume. But for the salt to start to flow under the weight of the overlying rocks they must be sufficiently (at least a few hundreds of metres) thick.

A typical morphological feature of diapir domes is the narrow circular syncline around them, which forms at those places where salt had flowed most intensively during the development of the injection core.

The diameter and shape of the salt core apparently depend on the depth of occurrence of the initial salt stratum and on the properties of the rocks through which the salt rises. In those cases when the salt pierces different rocks the shape of the core is irregular which, is expressed in the formation of overhangs at its sides, warping of the entire stock, etc. (see Fig. 48).

In the areas between the domes from which the salt was squeezed out, the thickness of the salt bed should be greatly reduced probably to the point of complete pinching out. This, however, has not been positively established because such areas are too deep-seated.

Injection folds may result from the squeezing of plastic material from beneath depressions under the load of the overlying rocks. These are thicker, and, therefore, heavier than the rocks of the adjacent uplifts, where cover rocks may be altogether absent. This process, however, may develop only when the specific gravity of the plastic series is less than that of the overlying rocks, or if there exists a reversed relief, that is where a higher relief corresponds to depressions (synclines) than to uplifts (anticlines). It seems in this way that the clay diapir injection folds of the Kerch Peninsula were formed. Two factors were at work there. First the water- and gas-saturated (owing to the activity of mud volcanoes) Maikop

series is not only very mobile, but also lighter than the overlying Miocene and Pliocene strata; second, the anticlinal arches of the Kerch Peninsula are eroded and gypsometrically lie below the surfaces of synclines.

Fig. 67 shows an example of an injection fold in the foothills of the French Alps (Provence) produced by the mechanism of squeezing of material by gravity from beneath depressions. Here the injection core is composed of gypsums and plastic thinly laminated marls. The core suffered intensive squeezing and was extruded onto

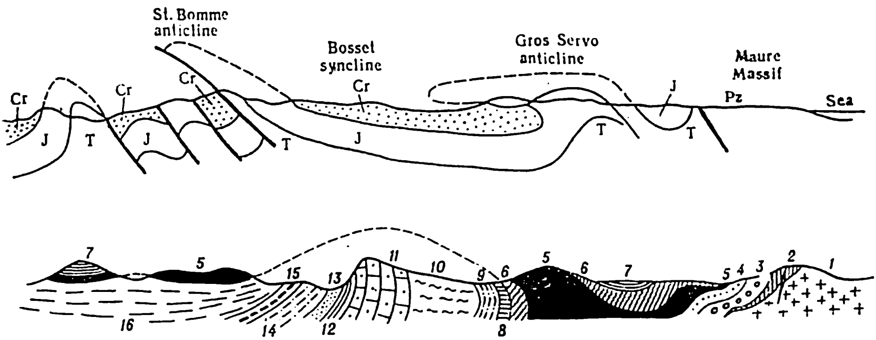


Fig. 67. Injection folds in Provence (France). Lower profile:
1—phyllites; *2*—Carbon; *3*—Permian; *4*—variegated sandstone; *5*—shell limestone; *6*—Keuper; *7*—Infralias; *8*—Lias; *9*—Bathonian; *10*—Upper Jurassic; *11*—Urganian; *12*—Aptian; *13*—Cenomanian; *14*—sandy Turonian; *15*—Turonian with rudists; *16*—Cenonian

the erosion surface of adjacent rocks as a tongue or, as it were, a small sheet overthrust, or fold-nappe, subsequently eroded away except for some outliers. This mass has a horizontal overthrust breadth of up to 5 km and carries on its "back" fragments of overlying Jurassic limestones.

Non-uniform load on a plastic series is often due to ruptures in the overlying rocks. In the ruptured zone the pressure on the plastic series decreases, allowing the plastic material to rise along the rupture, which results in the development of injection folds.

Injection folds of the gravity extrusion type may also develop even when the strata are horizontal due to irregularities of relief alone. If a plastic series is exposed on the floor of a gulley or valley, while at the sides it is overlain by other rocks, the plastic series at the floor may be squeezed out and form an injection core (Fig. 68). The injection folds in the thick coal seams of the Chelyabinsk basin have apparently formed under such conditions.

Lastly, injection folds may result from the pushing aside of plastic material from the arch of a growing uplift, as mentioned earlier on page 86. The material that had been pushed aside may concentrate

along the side of uplift, form an injection core and be squeezed upwards, particularly. if in this process it can exploit ruptures in the overlying series (Fig. 69).

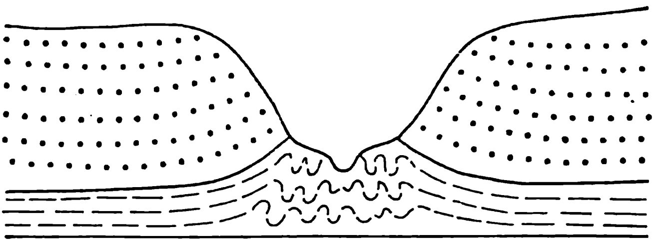


Fig. 68. Compression folds due to surface relief

In general, it should be noted that ruptures in the overlying series play a very important role in the formation of injection folds and

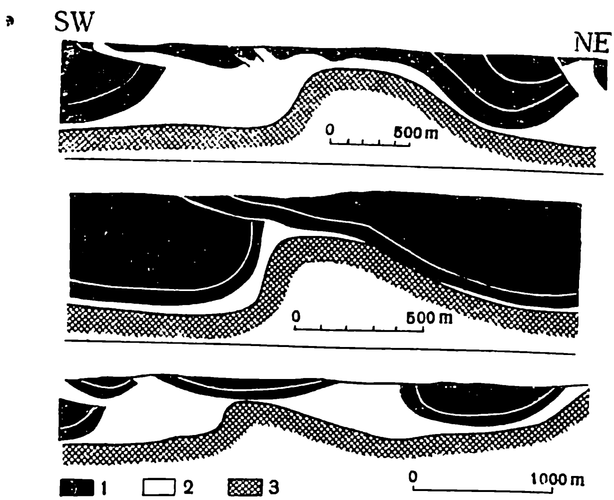


Fig. 69. Squeezing of plastic rocks from the arches of block folds. Iran (after O'Brien):

1—upper hard series (Middle Miocene—Pliocene); *2*—plastic series (Lower and Middle Miocene); *3*—lower hard series (Lower Cambrian—Lower Miocene)

often determine the pathways for the upward movement of plastic rocks and the shape of the injection core. This gives rise to dyke-like, circular, arcuate, etc., injection cores.

The development of injection folds was repeatedly modelled by Soviet and American researchers. Diapir gravity buoyancy domes were reproduced on two liquids: a bitumen, which simulated salt, and a heavy sugar syrup, which represented the country rocks. The density of the bitumen (0.95) was less than that of the syrup (1.1 to 1.2). A layer of syrup is poured over a bitumen layer. Before long a movement, clearly observable through the syrup, begins on the surface of the bitumen, on which small knolls are formed. These knolls grow into short columns which keep rising and assume a shape closely resembling that of salt diapirs in nature. Modelling helped to establish certain features of the development of domes. In particular, it was demonstrated that the growth of the diapirs was most intensive where the thickness of the "salt" (represented in the model by bitumen) was greater.

CHAPTER V

Disruptive Dislocation of Rocks

Morphology

Disruptive dislocations of rocks are divided into two principal groups: dislocations without displacement and dislocations with displacement, also called diaclases and paraclases respectively, though in our country these terms are no longer used. We shall refer to them as *joints* and *faults*. It should be noted that the concept of joints as fractures without displacement is a conventional one, since fractures without any displacement of the walls do not exist. In nature some, however slight, displacement is always observed, either in the form of separation of the sides or their slippage relatively to each other. But these displacements are so small as compared with the length of the joints that they may be ignored.

18. JOINTS

Joints are very common in the earth's crust. They occur in nearly all rocks, except the most loose ones or those that readily soak water and begin to flow due to which the joints are closed.

Joints differ among themselves in the degree of their gaping, size, form, position in space and relatively to structural tectonic elements (folds, internal structure of an intrusive rock, etc.).

Depending on their gape joints are called *latent*, *closed* and *open*. Latent joints are invisible in their natural state due to their extreme thinness. They betray their presence when a rock is broken, whereupon it splits along these joints. Closed joints are visible to the naked eye but have no noticeable cavity. Open joints have a visible cavity.

Joints may be small and confined to a single stratum, or large and cutting across a series of strata. Most common are joints of the first kind. Usually they traverse the strata at right angles to the bedding. Systems of such small joints are referred to as *jointing* (Fig. 70).

Large or intersecting joints split series of strata and sometimes extend for many kilometres along the strike and the dip. Some joints or fractures extend for tens of kilometres along the surface of the earth. In most cases, however, major fractures are accompanied by displacement.

Joints may be straight or curved and their sides may be smooth, slickenslided, or jagged or serrated. As in describing the attitude of bedded rocks, the orientation of joints may be designated by their strike and dip.

Joints may be perpendicular to the bedding, parallel to the bedding, or oblique to the bedding. Perpendicular, oblique joints may be parallel to the dip or to the strike of a stratum, or diagonal to it. The orientation of joints relatively to the strike of linear folds,

Fig. 70. A jointing exposure

may be longitudinal, transverse or oblique. In the case of domed forms one may speak of radial and concentric joints.

A somewhat different classification of joints is applied to massive rocks characterized by internal orientation of minerals. In this case joints may be parallel or normal to the linearly-oriented structure and are referred to as longitudinal and transverse, respectively. Relatively to the outer perimeter of a mass the joints may be marginal, running parallel to its edges; radial, when directed at right angles to the edges, or surficial reproducing in a subdued way the surface relief (see Fig. 20).

Very interesting are the dependences governing the mutual arrangement of joints. Generally they occur in systems. A system consists of joints that are parallel in strike and dip. Sometimes, however, the term system is applied to a different arrangement of joints. For example, in domes systems of radial and concentric joints may be distinguished.

As a rule systems cross each other at different angles. In the same stratum or in the same rock mass the frequency of joints in each (parallel) series, that is the number of joints per unit length in a direction normal to the orientation of the joints, is usually sustained over large areas. Different systems are usually characterized by different frequency of jointing.

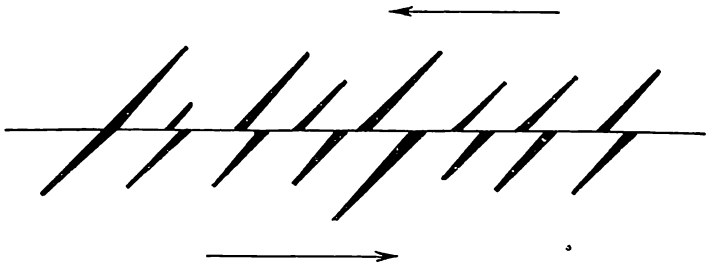


Fig. 71. Feathering joints. Arrows show the direction of the couple of forces that caused the shear deformation

Other conditions being roughly equal, the frequency of joints varies with the nature of the rocks and their thickness.

Intersecting joints divide rocks into blocks of different shape and size. In bedded rocks one of the systems always lies in the bedding plane. The shape of such blocks depends on the direction of the

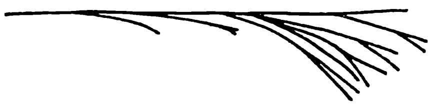


Fig. 72. Branching of tectonic fissures (horse-tail structure)

systems of joints and the attitude of the strata, the frequency of jointing and the thickness of strata.

Depending on the shape and dimensions of the joint blocks are referred to as cubic, block, platy, etc., though there is no general agreement on the meaning of these terms.

Systems of echeloned joints are common and may be right-hand or left-hand. In a left-hand set of echeloned joints they recede to the left from an observer looking along the line of their development. In the right-hand system the situation is reversed.

Often echeloned joints abut a fracture extending along the strike of the entire system of joints. In this case they are known as *feather joints* (Fig. 71). From what has been said above on the formation of joints in shear dislocations, it follows that feather joints are tension

joints, whereas the longitudinal fracture is a shear joint parallel to the couple of forces that caused the shift.

Occasionally branching joints are observed. The so-called *horsetail* structure of a bundle of branching fractures is shown in Fig. 72.

19. FAULTS

Several kinds of disruptive displacements are recognized, depending on the attitude of the fracture and the direction of the movement of its walls.

The two principal kinds of disruptive displacements are ruptures and faults.

Ruptures

Ruptures are displacements in which the walls are simply torn apart without being vertically or horizontally displaced relatively to one another. This occurs mostly when the strata are broken into blocks and also in the course of jointing.

A rupture may be characterized by the magnitude of its gap. Gaping ruptures are shortlived even near the surface because usually they are closed by the creep of the wall rocks or filled by extraneous magmatic mineral matter or material brought by water, or by plastic material squeezed from adjacent beds. Usually only small (not bigger than one or two centimetres), mostly intrastratal, ruptures remain open. The walls of ruptures filled by mineral matter sometimes may be hundreds of metres apart but most often the width of the opening does not exceed a few tens of metres.

Rupturing is often associated with displacement of the walls parallel to the rupture plane, for example, with normal faults, which as it will be seen later, are frequently related to extension of the earth's crust.

Faults

Faults are displacements in which the walls of fractures are displaced parallel to the fracture. The following types of faults are distinguished: *strike-slip faults*, *normal faults*, *reverse faults* and *overthrusts*. All have certain morphological features in common which are best considered together.

The fracture along which faulting occurs is called the *fault plane*. The rocks that adjoin the fault plane are called the *walls* of the fault. When a fault plane is inclined the displaced part that is above it is called the *hanging wall* and the part that lies beneath the fault plane is called the *foot wall*. A fault is characterized by the orientation of the fault plane relatively to the folding and also by the direction of displacement and its throw. Faulting along an

oblique fracture (i.e., one which involves displacement both along the strike and the dip) is represented diagrammatically in Fig. 73*a*. The distance *ab* represents the total amount of throw or displacement. It may be expressed by the azimuth of the strike and the dip angle and also by the angle *bac* between the line of displacement and the strike of the fault plane.

Distance *af* is the horizontal slip or displacement. Distance $ac=db$ is the displacement along the strike or the *strike-slip* and distance $cb=ad$ represents the displacement along the dip or the *dip-slip*.

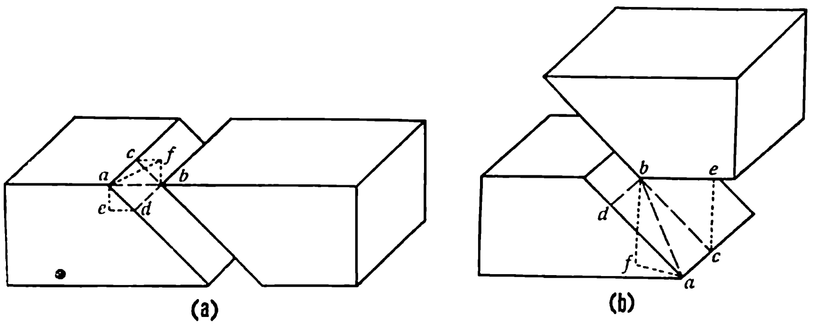


Fig. 73. Throw of gaping fault (*a*); throw of a fault with doubling of strata (*b*); the letter symbols are explained in the text

The vertical line *ae-fb* represents the vertical throw or vertical displacement. The horizontal displacement in the direction normal to the strike of the fault plane represents the opening between the walls of the fracture. Fig. 73*b* gives another fault for which it is possible to determine the amount of throw (*ab*), angle of inclination (*cab*), horizontal shift (*af*), vertical displacement ($ce=fb$), strike-slip ($ac=db$), dip-slip ($ad=cb$). In this case however, instead of speaking of opening, one should speak of convergence.

Since in most cases faulting brings into contact strata of different ages, one may speak of the *stratigraphic throw* of a fault. This cannot be expressed quantitatively but only indicates the stratigraphic interval that is missing or repeated in the section as a result of the given faulting. This interval depends not only on the direction and the geometrical throw of the fault, but also on the position of the strata and the thickness of individual stratigraphic subdivisions.

From the practical point of view faulting may be important when it results in doubling of strata, or in a hiatus in their sequence. Doubling occurs when a perpendicular erected to the stratum meets the same stratum at its extension beyond the fault plane (Fig. 73*b*). A hiatus occurs when the stratum will not be found under the same conditions (Fig. 73*a*).

In working bedded deposits (coal, for example) doubling is a favourable factor since it increases the mineral reserves of the given area. Conversely, a hiatus reduces the available reserves.

It should always be borne in mind that, as a rule, the direction and throw of faulting are liable to change both along the strike and dip of the fault plane and that all faults diminish in scope and die out altogether at a certain distance.

We shall now proceed to discuss the different types of faults from the morphological point of view.

Strike-slip faults. A strike-slip fault is a displacement along the strike of the fault plane. The fault plane may be vertical or inclined at any angle. Depending on the direction of the displacement, right-hand and left-hand strike-slip faults are recognized. If one looks at a strike-slip fault at right angles to the fault plane, in the case of a right-hand shift the side that is farthest from the observer is displaced to the right. In a left-hand shift, under the same conditions the displacement is to the left.

In unstratified rocks a strike-slip fault is characterized by right-hand or left-hand displacement, its throw and fault plane orientation. In stratified rocks with inclined bedding longitudinal, oblique and transverse strike-slip faults may be recognized. A longitudinal strike-slip fault does not result in any observable displacement either in plan or in cross-section. There are also distinguished concordant strike-slip fault in which the fault plane dips with the beds and discordant strike-slip faults in which the fault plane dips against the dip of the beds.

A *normal fault* is a dislocation parallel to the fault plane in which the hanging wall has been depressed relatively to the foot wall.

Depending on the orientation of the fault plane relatively to the strike of the beds or folds, normal faults may be *longitudinal*, *transverse* or *oblique*. If the fault plane dips with the bedding it is said to be concordant, and one that dips against the bedding is referred to as discordant.

Because of the fact that the throw of a fault changes along the strike of a fault with a vertical fault plane, the fault may not only fade out completely but the foot wall may become the hanging wall and vice versa. This is referred to as a *hinge fault*.

In *reverse faults* the fault plane usually dips at a high angle and the foot wall is raised relatively to the hanging wall. Depending on the orientation of the fault plane relatively to the strata several types of reverse faults, similar to those of normal faults, are recognized.

An *overthrust* (or *thrust fault*), like the reverse fault is characterized by an inclined fault plane and upward displacement of the foot wall, but the fault plane dips at a lower (less than 60°) angle.

However, the distinction between reverse faults and thrusts is a purely conventional one and in many cases cannot be positively made. Some geologists prefer to describe reverse faults as steep thrusts. Another (kinematic) basis for distinguishing reverse faults from thrusts will be considered later.

In major thrusts the fault plane may be undulating. Locally this plane may have a reverse incline; in such sections the thrust is said to be plunging (Fig. 74). In other sections it may be overturned

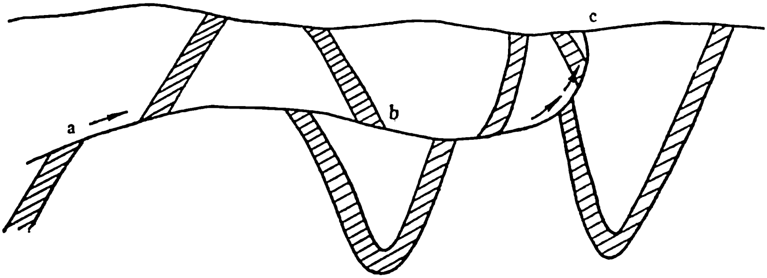


Fig. 74. Thrust fault:
a—normal; *b*—plunging; *c*—reverse

(see Fig. 74). In most cases the undulation of the thrust plane is due to later dislocations and, therefore, is not related to the development of the thrust as such.

Thrusts characterized by great horizontal displacement (from a few to tens of kilometres) over a low-dipping undulating surface are referred to as overthrusts or nappes. In overthrusts the stratigraphic sequence is often abnormal, older sheets being thrust over younger ones. Conversely younger sheets may be thrust over older strata. The rocks of overthrust sheets are referred to as *allochthonous* (foreign), while those that have remained in place are called *autochthonous* (native). Tectonic overthrusts are characterized by different structures of the allochthon and autochthon. Both are usually crumpled into folds, but the character and the dimensions of the folds as well as their location are different. Intensely crumpled rocks of the overthrust sheet may overlie undisturbed allochthonous strata; in other instances the allochthon may be characterized by more quiet bedding than autochthon. Allochthon may consist of several sheets overthrust upon each other. In this case the sequence of the layers in the section may be most irregular.

Where erosion has cut through the allochthonous strata, so-called geologic windows or semi-windows are formed through which these strata are exposed. In other cases erosion leaves only remnants of the

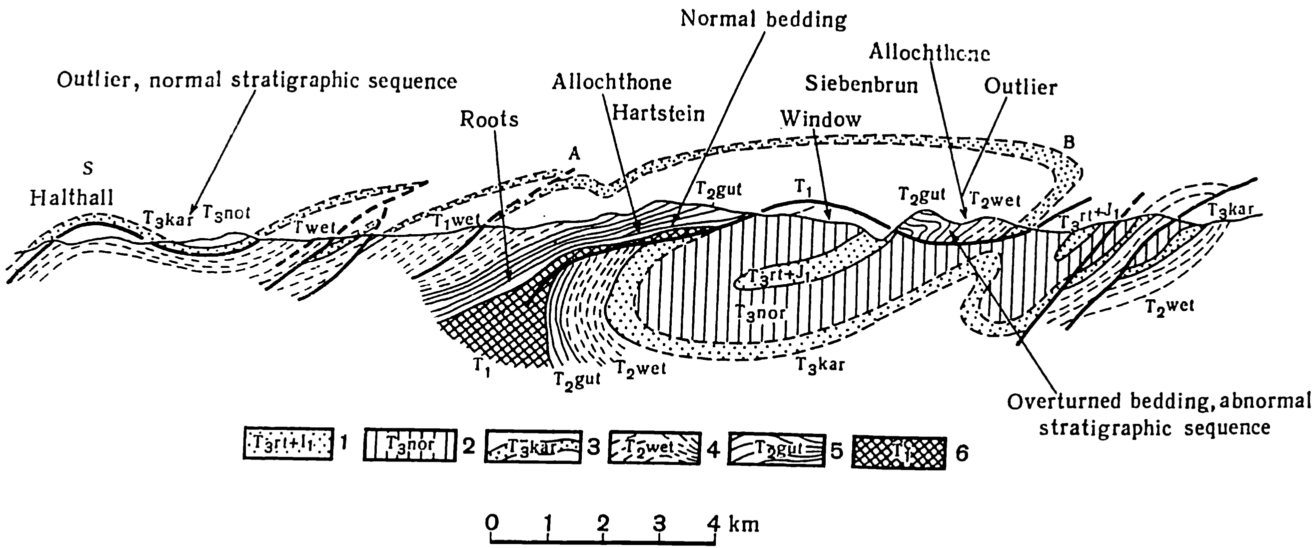


Fig. 75. Small overthrust. Annaberg recumbent fold (Austria).

1-6—stages from Rhaetian to Lower Triassic from top to bottom (Lower Triassic is represented by very plastic rocks)

overthrust sheets, called outliers, which are detached from their roots, i.e., the places where the sequence of the strata is tectonically normal and from which their movement began (Fig. 75).

Complex faults. The above classification of faults covers those displacements that occur either only along the strike or only along the dip of the fault plane. In nature such simple displacements are rare. It is true that in many cases it is possible to ignore a slight deviation of the displacement from a certain "pure" direction. However, most faults observed in nature are of the complex type, that is, involve displacement both along the dip (vertical) and the strike horizontal). The resultant is usually inclined.

CHAPTER VI

Natural Conditions of the Development of Dislocations

20. DIVISION OF ROCKS BY JOINTS AND FRACTURES

General Joints

All rocks, regardless of their mode of occurrence are divided by general joints.

The cause of the development of general joints present a problem for which no adequate explanation is yet forthcoming. Presumably, most of these joints develop as a result of contraction of rocks on their dessication (sedimentary rocks) or cooling (igneous rocks). As known, when marine sediments are consolidated into rocks they lose a tremendous amount of moisture and greatly diminish in volume. Igneous rocks, too, contract on cooling and crystallization.

Had rocks occurred as bodies freely hanging in space they could have contracted without cracking. But rocks occur as bodies (strata, masses) that overlie and are in turn overlain by other rocks. Since contraction causes rock particles to move, at the contact with other rock bodies, such movement is inevitably accompanied by friction. Since friction acts against the forces that cause the rock to contract, it constitutes a tensional factor. Due to this tensional action the rock body does not shrink as a single whole but splits into numerous small blocks each of which contracts separately. The elementary unit volumes (blocks) are separated from each other by tension joints. Each joint usually cuts through a single bed and does not pass into other beds. Most of them in stratified rocks are vertical relatively to the bedding planes. In inclined strata they are most often oriented along the strike, dip and also diagonally. The causes of such orientation and also of the occasional deviations from it are not known. A most general explanation is that the tectonic movements that caused the tilting of the strata produce a certain anisotropy in the properties of rocks, which in turn controls the directions most favourable to the development of general joints.

Apparently it is this primary anisotropy that explains the localization of joints in igneous rocks in which they are mostly oriented either normal or parallel to linear texture. In the absence of oriented textures, the joints are either irregularly oriented or tend to divide the rock into hammock, ovoid or spherical bodies.

Columnar or prismatic structures are characteristic of congealed effusive flows.

A well-defined relationship is observed between the frequency of general joints and the mechanical properties of a rock. Other conditions being equal, in harder rocks and in thicker layers, joints are less frequent, whereas thinner layers composed of weaker rocks are divided by more numerous cracks. We had already dealt with a similar phenomenon in examining the process of formation of lenses and blocks on the limbs of folds under the action of compressive forces.

Some scientists believe that the regular arrangement of general joints, which is most surprising in the case of nearly horizontal platform strata, bears evidence against their non-tectonic origin. For example, on the Russian Platform the systems of general joints are similarly oriented over tremendous areas. A definite relationship is also observed between the direction of jointing and the orientation of some very gentle folds of the discontinuous type.

Indeed, the most probable manifestation of non-tectonic division of rocks due to contraction would be the splitting of the rocks into numerous polygonal prisms, like those observed in a cooling lava flow. From this point of view it is much more difficult to explain rectangular jointing. On the other hand, the mechanism linking jointing to tectonics is not known. The negligible tilting of strata observed on platforms can hardly help to explain the orientation of the joints.

A more promising approach perhaps would be to relate general jointing to the fundamental division of the earth's crust to which we shall return later.

Tectonic Fractures

Tectonic fractures are produced by tension, compression and shear which develop in the earth's crust under certain tectonic conditions. For simplicity of classification let us assume that tension and compression act in the horizontal direction, while shear may act either in the horizontal or vertical directions.

First let us consider fractures caused by horizontal extension. We have noted earlier that either tension or shear fractures may develop under these conditions. Figure 64 represents an amphibolite bed in gneisses which was subjected to extension and as a result was divided into rectangular blocks. Inasmuch as the tensile stresses acted along the plane of the bed, the fractures are of tensional origin. Often, however, a bed is divided by oblique fractures (approximately at an angle of 45° to the axis of extension). Furthermore, were blocks brought together the thickness of the bed would not be uniform. The

shape of the bed will clearly indicate that before division into blocks it was subjected to boudinage with the development of pinches or necks. This is a perfect case of division of a bed in the process of its extension by shear, or to be more specific, by viscous shear, inasmuch as it was preceded by plastic deformation.

Whether tension fractures or shear fractures develop in each specific case depends on the plasticity of the rocks and the rate of deformation. Rapid deformation in brittle rocks tends to produce tension fractures, whereas slow extension of plastic rocks results in shear fractures.

Now we shall consider the conditions under which rocks may be subjected to extension.

Let us conceive a stratified rock series in which bed *2* of low plasticity is confined between two more plastic layers *1* and *3*. The series is subjected to compression normal to the bedding with uniform distribution of stresses. The stresses exceed the yield points of beds *1* and *3* which suffer plastic deformation by crushing. In this process they grow thinner but are lengthened and spread out.

As to the bed in between whose yield point is higher it is either still in the stage of elastic deformation or, due to its greater viscosity, spreads more slowly than the other beds. In either case layers *1* and *3* "flow" past bed *2*, whose deformation is slower. As a result the surface planes of the middle bed are subjected to the action of tangential friction forces of a tensile nature, due to which bed *2* is elongated. As we have said earlier, such conditions often develop on the limbs of folds of general crumpling.

Extension may also develop in the case when a bedded rock sequence is upwarped under the action of tectonic forces (transverse bending). Inasmuch as at the edges of the deformed zones the beds stay in place, the surface of the uparched beds proves to be greater than the original horizontal surface, and as a result the beds are extended even in the absence of more or less plastic rocks. A series composed of rocks of low plasticity will also crack at the most sharply bent part if the tensile stress becomes great enough.

The two cases have one feature in common, the tensile stresses are applied to the entire surface of the bed and operate at all its points. This is very important. Let us imagine that the bed is extended by forces applied at two points (the "ends"). Then as long as the bed remains whole and under elastic extension, strain is observed along the entire length of the bed, and any two points on its surface move apart. But as soon as the bed is sundered by a fracture, the picture changes abruptly. The development of the fracture relieves the tensile stresses in the bed, and, if the forces continue to operate, the two separated parts of the bed will move apart. No other fractures can develop in the bed in this case.

A different picture is observed when tensile stresses are applied to the entire surface of the bed. Then a fracture that has developed at a certain point does not relieve the bed completely of tensile stresses and they continue to act in the sundered blocks. Therefore, if the stresses are great enough, new fractures may develop in the blocks. Their number will be the greater the greater are the tensile stresses and the lower the strength of the bed, which in turn depends on its lithological composition and thickness. Other conditions being equal, fractures are more widely spaced in stronger and thicker beds than in thin and weak layers.

An example of the division of a layer into blocks are the so-called "ladder veins" in dykes in which tension cracks, produced by the dragging apart of harder dykes confined between two more plastic layers, are filled with mineral matter (Fig. 76).

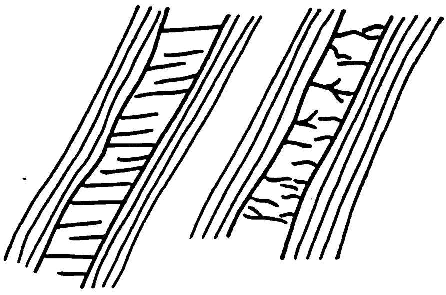


Fig. 76. Ladder veins. Quartz veins amidst clays suffered extension under compression normal to the layers. Transverse tension cracks developed in quartz as a result

Variations in the spacing of fractures in one and the same bed indicate the magnitude of deformation suffered by its different parts. All other conditions being equal, more closely spaced fractures bear evidence to greater deformation in the direction of the extension responsible for the given fracturing.

The diagram given in Fig. 63 shows limestone beds confined between more plastic marls. The limestone beds were subjected to extension and dragged apart into blocks along shear joints in two directions—along the dip and along the strike. However, the dimensions of the blocks in these two directions are not the same. Along the dip their cross-section is much smaller than along the strike. This shows that the tensile strain along the dip was several times greater than along the strike.

It is evident that upon compression of the limbs of the fold the plastic marl beds spread out both along the dip and the strike, and that the spread along the dip was much more intensive.

If tension fractures on tectonic uplifts developed in the course of uparching, their arrangement is controlled by the shape of the uplift. A round uplift is characterized by systems of radial and concentric cracks (Fig. 77). Together they produce what is sometimes referred to as "turtle structure" (Fig. 78). Such fracturing corresponds to a more or less uniform extension of the layers caused by the elevation of a round dome.

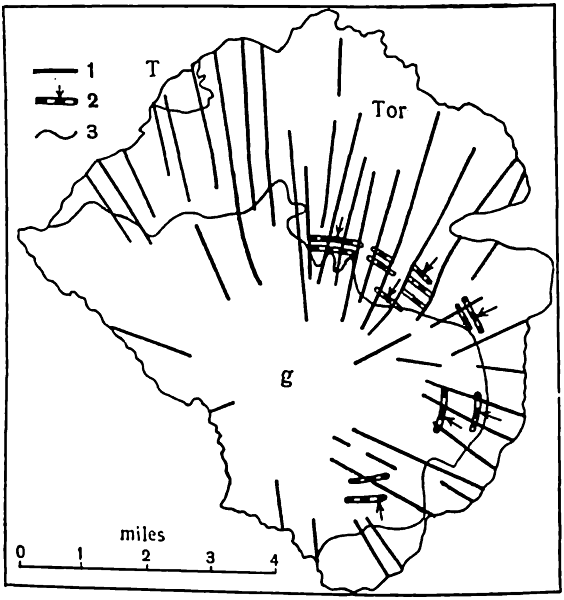


Fig. 77. Arrangement of cracks on a round tectonic uplift. Radial and concentric fissures filled by dykes on the Rum Island, Scotland:

1—radial dykes; *2*—concentric dykes; *3*—boundaries of rocks; T—Triassic; Tor—Torridonian sandstones; g—intrusive rocks (granite, gabbro, etc.)

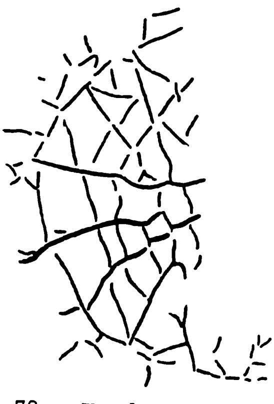


Fig. 78. "Turtle-back structure" (schematic)

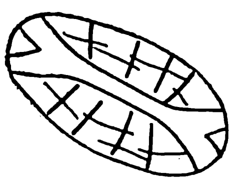


Fig. 79. Tectonic fissures on an oval (brachyanticlinal) uplift (schematic)

If the uplift is oval (brachyanticline), both longitudinal and transverse cracks are observed, the former having developed earlier (Figs. 79 and 80). This is linked with the fact that the strata are bent

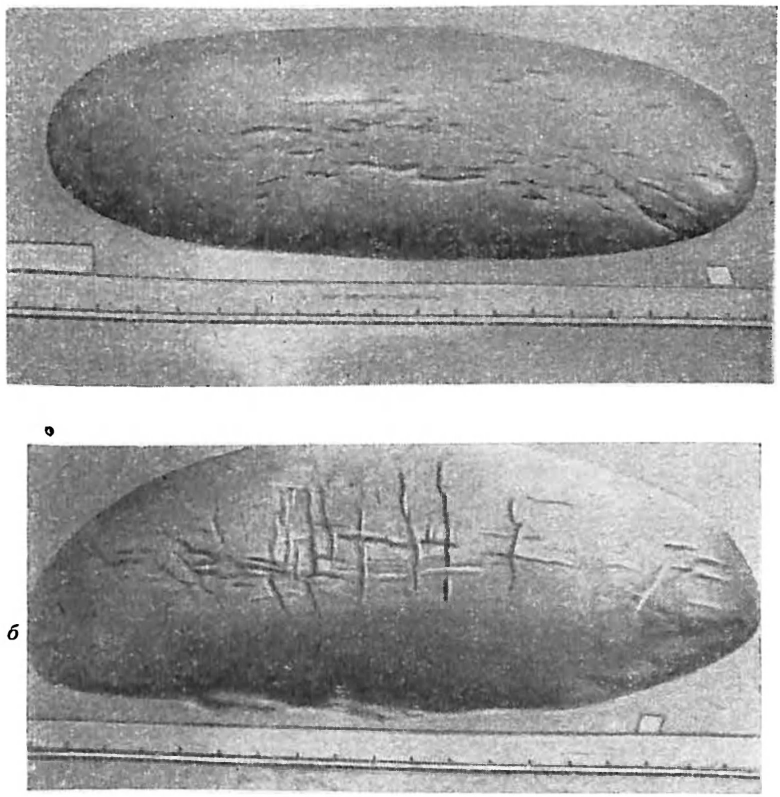


Fig. 80. Fissures obtained on a clay model of a brachyanticline: *a*—first stage—development of longitudinal cracks; *b*—second stage—development of transverse cracks (after M. Gzovsky)

more strongly transversely to the fold than along it, and therefore a degree of tensile stresses great enough to cause cracking develop earlier in the transverse direction.

Longitudinal cracks develop most intensively in the arched part of an uplift. Periclines are usually characterized by radial cracks (see Fig. 79). On some brachyanticlinal uplifts either radial or transverse cracks alone may be observed (Fig. 81).

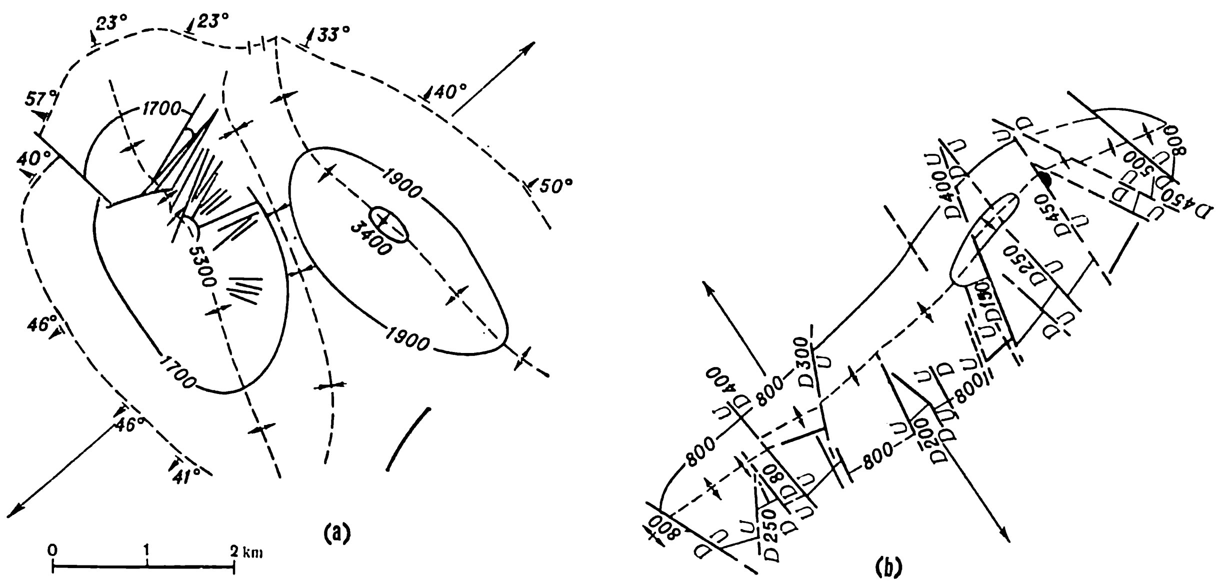


Fig. 81. Brachyanticlines dislocated by fissures:
a—radial; *b*—transverse

Tectonic compression usually causes the development of shear fractures. Practically in every case they immediately pass into faults, whose mechanism will be considered in the next chapter.

Shear causes the development of both shear fractures (parallel and perpendicular to the acting couple of forces) and tension fractures. The latter in the typical case are expressed as echeloned fractures, which are widespread in nature. They may be from a few centimetres to several kilometres long. Shear deformations develop in the earth's crust under diverse conditions, when one block of the earth's crust is displaced relatively to another in any direction (for example, when one block is uplifted or slips horizontally relatively to another or in the course of relative slipping of strata subjected to folding).

21. DEVELOPMENT OF FAULTS

Faults develop as a result of continued displacement of the sides of fractures. We shall consider various cases of the development of faults under the action of horizontal tension, horizontal compression, or shear stress.

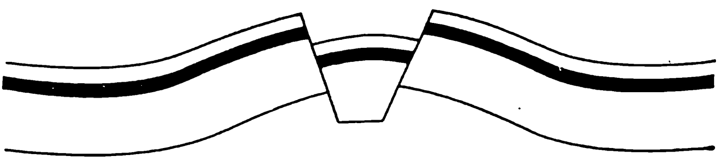


Fig. 82. Development of a fault trough on an uplift under extension

If tensile stresses continue to act after the development of a fracture, the walls of the fracture will move apart, resulting in what is called a rupture. This is most characteristic of the splitting of strata into blocks and their further detachment.

Development of openings in the earth's crust is intensified by gravity which causes downward displacement of separate blocks. This results in typical normal tension faults.

Tension faults develop most frequently at the crests of tectonic uplifts. The pattern of development of such faults is as follows: a block of the earth's crust, previously bounded by fractures sinks into the opening formed due to extension and wedges into it (Figs. 82 and 83). Such "wedging" may occur in the case of a single fault, if a corresponding plastic bending of strata is observed on the other side of the downfaulted block. More often, however, faults occur in groups, bounding the sunken blocks on different sides.

The most typical combination of faults is the *fault trough* or *graben*, that is a combination of two faults, the block between which has sunk. In other cases fault troughs are complicated by minor faults. In some instances they complicate the walls of the trough, forming

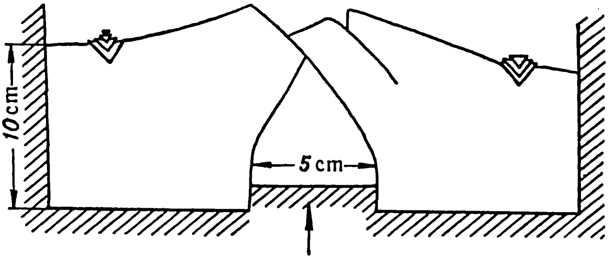


Fig. 83. Model of a fault trough obtained by bending paraffin by means of a die

a series of step-like faults (step faults). Some blocks may be upthrust against the background of a complex fault trough to form so-called *horsts* (Fig. 84).

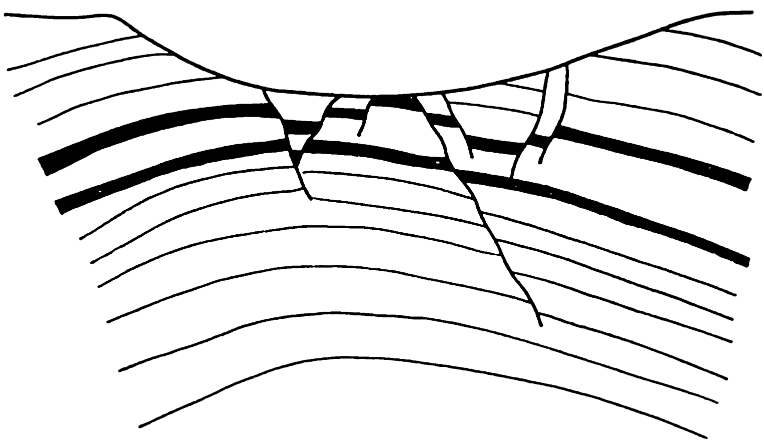


Fig. 84. Complex horst (on the arch of a tectonic uplift) (Apsheron Peninsula)

Thus, within boundaries of a region of general subsidence (complex fault trough) there may be observed a very complicated system of numerous faults with different directions of displacement, whose common characteristic is that they are associated with extension of the earth's crust.

However, the conditions under which faults and combinations thereof such as fault troughs and horsts occur are not restricted to that. A horst often develops by the uplift of a block of the earth's crust between two faults. In this case the fault trough may be subordinate to horsts, representing segments of the earth's crust that

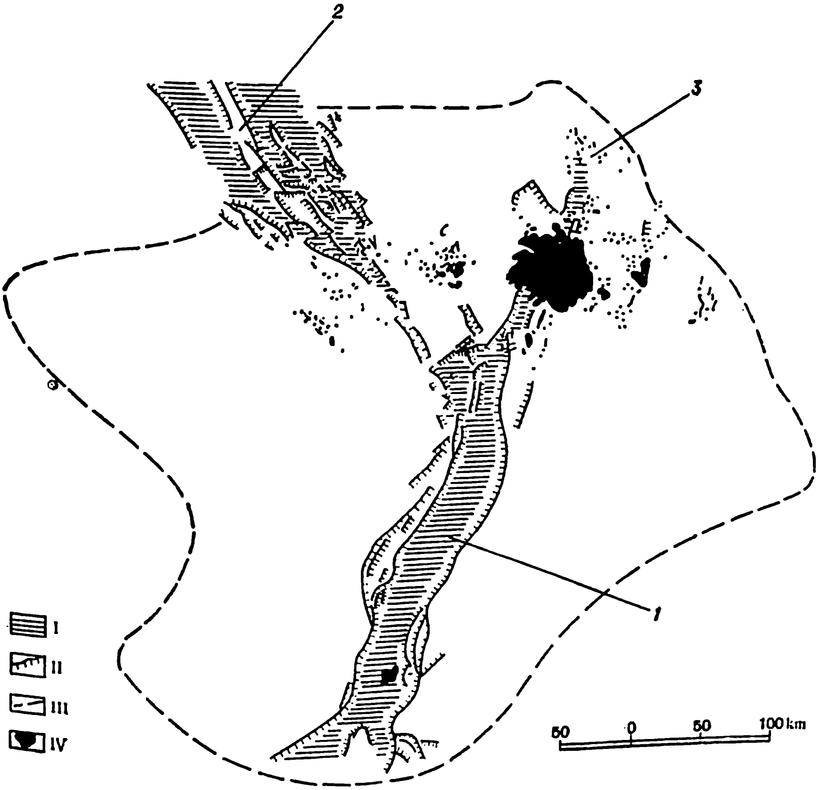


Fig. 85. Schematic representation of the Rhein system of fault troughs.
I—fault troughs: *1*— Upper Rhein; *2*—Lower Rhein; *3*—Hessen; *II*—faults; *III*—contour of the Rhein anteclise; *IV*—regions of Tertiary and Quaternary volcanism

have lagged behind in general upward movement. Similarly, fault troughs often develop regardless of extension on tectonic uplifts, but as a result of independent downward movement of a block under the action of certain, still poorly understood deep-seated processes. Often horsts and fault troughs alternate, as a result of which a tract of the earth's crust looks like a keyboard whose keys are alternately depressed or raised. Dislocations of this kind should be viewed in connection with faults caused by shear.

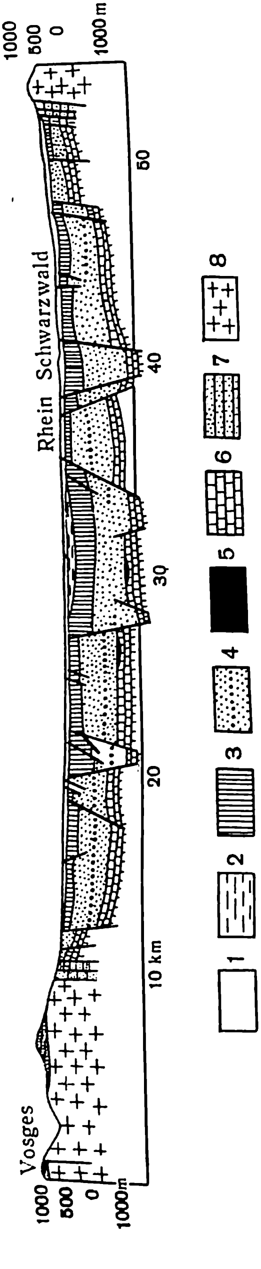


Fig. 86. Profile through the Rhein graben:
1—Quaternary deposits; 2—Miocene; 3—Upper Oligocene; 4—Middle Oligocene; 5—Eocene; 6—Jurassic; 7—Triassic; 8—folded basement

Faults may develop as a result of either tension or shear fractures, which is expressed in the steepness of the fault plane (almost vertical in the case of a tension fracture and inclined in the case of shear).

Examples of major faults developing under the conditions of extension of the earth's crust at the arches of upwarps are the grabens of the Baikal trough, the Rheingraben, the East African system of rift valleys, etc.

The Baikal system of faults consists of a large number of fault troughs tending generally from southwest to northeast along the axis of the great Baikal arch. This is an anteclise that had experienced prolonged tectonic upwarping in the Paleozoic and Mesozoic times. Longitudinal fractures began to develop within this anteclise in the Tertiary and Lower Quaternary. The downthrow of the biggest fault trough, that of L. Baikal proper, is six kilometres.

The Rheingraben has developed as a result of extension and fracturing of the earth's crust in the course of the elevation of an anteclise that formerly (before the beginning of the Tertiary period) encompassed the Vosges and Schwarzwald, now separated by the Rheingraben. The subsidence that produced the fault trough occurred in the Middle Tertiary and its throw exceeded 1 km (Figs. 85 and 86).

The greatest fault troughs or rift valleys are observed in East Africa. Two fault systems are known here: the Great Eastern Rift extending from the mouth of the Zambesi River in the south to Ethiopia in the north, that is along entire East Africa, and the rift valley extending between Arabia and Africa, the centre of which

is in the Red Sea. The two systems join in Ethiopia and form a great complex of faults extending meridionally for 6,500 km. Both systems of rift valleys have developed against the background of vast upliftings or anteclises (Fig. 87) in the Tertiary period.

Smaller faults associated with the subsidence of crests of uplifts are most characteristic of the upper rock series in diapir folds which is subjected to intensive extension during the elevation of the diapir core (Fig. 88).

In all the cases described above the location of the faults, big or small, on uplifts is controlled by the same dependences that were recognized in the case of tension fractures. Oval uplifts are typically characterized by the development of a central longitudinal fault trough which branches out at the periclines where the fractures fan out. As evident from the accompanying illustrations this is characteristic of the Upper Rheingraben, the Red Sea fault troughs, and also of fault troughs that develop on the domes of diapir uplifts where transverse folds may develop as well. On round uplifts characterized by "turtle" fracturing the earth's crust is often split into an intricate mosaic of downthrust and uplifted blocks.

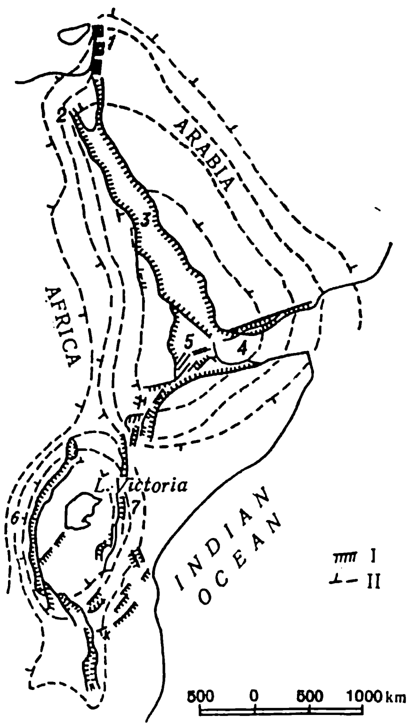
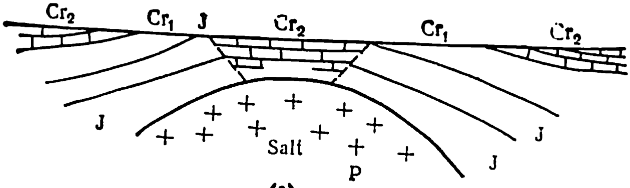


Fig. 87. African system of rift valleys: *I*—faults; *II*—isohypses of uplifts (anteclises). Rift valleys: *1*—Dead Sea; *2*—Suez; *3*—Red Sea; *4*—Gulf of Aden; *5*—Abyssinian; *6*—Western arc; *7*—Eastern arc

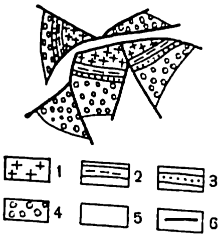
An exception is the East-African system of fault troughs or rift valleys. It will be seen from Fig. 87 that these rift valleys form two arcs: the Eastern, east of L. Victoria, and the Western arc which comprises the troughs of the lakes Rudolph, Albert, Tanganyika and Nyasa. Several troughs of smaller lakes are located within the Eastern arc. The two arcs have formed instead of a single median fault trough due to the non-uniformity of the structure of the earth's crust, specifically because of the great granite batholith which occurs in the region of L. Victoria and is surrounded by series of metamorphic and crystalline schists, at the boundaries of which the rift valleys have developed. Clearly, these boundaries offered less resistance to

extension than the axial zone of the anteclise. This shows that in some cases mechanical non-uniformities in the uplifted segment of the earth's crust may result in deviation from the usual arrangement of faults on uplifts.

It was stated earlier that the throw of a fault varies along the strike and that every fault terminates at a certain point. In the zone of fading the fault usually branches out and the displacement along every particular fault becomes progressively smaller. Thus, along



(a)



(b)

Fig. 88. Faults on diapir salt domes:

a—middle fault trough in the upper complex (Emba district); *b*—schematic geologic map of the Shunat diapir dome in the Emba district; *1*—Jurassic; *2, 3, 4*—Lower Cretaceous; *5*—Upper Cretaceous and Tertiary; *6*—fractures. Clearly visible are the middle fault trough and radial fractures

the strike a fault gradually passes into plastic deformation, most often represented by a flexure, which also fades out eventually (Fig. 89).

In plan faults appear in most diverse combinations, cutting across or abutting each other at various angles. Occasionally faults bound downcast blocks ("subsidence ovals"). The fading of faults along the strike results in various unilateral fault troughs and horsts and also cradle-shaped fault troughs, bridge-shaped horsts and also hinge faults mentioned earlier (Fig. 90).

As noted previously, *horizontal longitudinal compression* usually results in general crumpling. It is accompanied by a complex of faults which exhibit a regular relationship to the folding mechanism.

For example, it is known that when strata are crumpled into complete folds their material is squeezed into the crests. Such compression is accompanied by boudinage and dragging of the strata along shear or tension joints. The crushing of strata on the limbs normal to the bedding and their spreading parallel to the bedding are often accompanied by the development of a more complex combi-

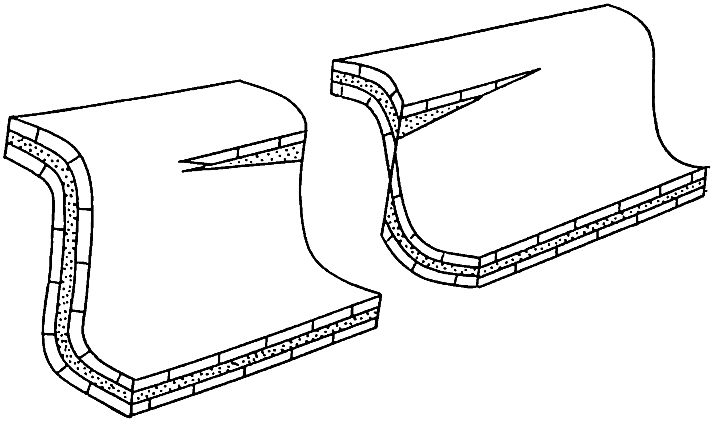


Fig. 89. A fault passing into a flexure

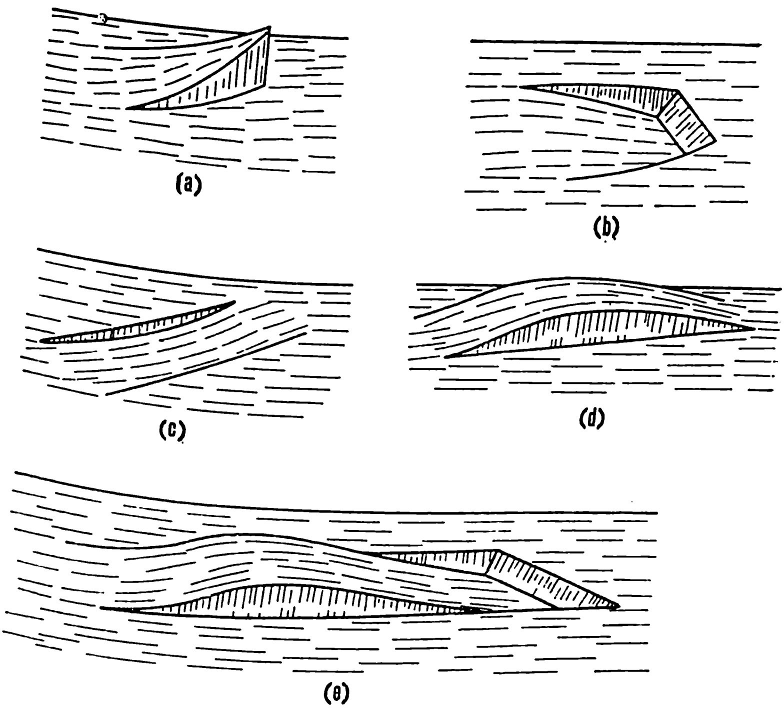


Fig. 90. Different kinds of fault structures:
a—unilateral horst; *b*—unilateral fault trough; *c*—cradle-type fault trough; *d*—bridge-type horst; *e*—hinge faults

nation of tension fractures, the sum of whose displacements is expressed in the extension of the stratum parallel to itself (Fig. 91). The movement of the material from the limbs to the curves may also be accompanied by fracturing which bears evidence to injection of material. An example is the occasional thrusting of limbs over

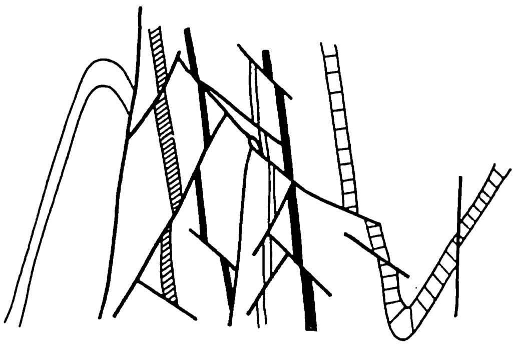


Fig. 91. Extension of a limb of a fold expressed in a system of faults. Kuznetsk basin (after A. Belitsky)

the crest of an anticlinal fold (Fig. 92). These overthrusts record the movement of the material into the crest of the anticline. The movement of material towards synclinal curves may result in similar fractures. Non-uniform squeezing on the limbs may cause the lower

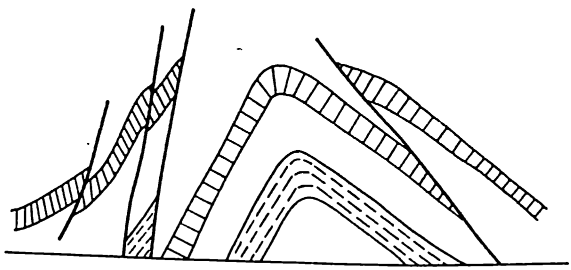


Fig. 92. Thrusting of limbs over the crest of an anticlinal fold (after S. Rumyantsev)

layers to "overrun" the upper ones which produces local small areas of injection or piling up of the material. Morphologically, this may be recorded in a system of small fractures on the limbs with repetition of strata.

General crumpling is accompanied by overthrusts and shifts of a corresponding scale. Most common is the combination of complete

folds with overthrusts. Thus, overthrusts should be regarded as a further development of deformation by folding.

Since a fold develops non-uniformly, by comparing its different parts it is possible to trace all the phases by which a fold passes into an overthrust. Usually the uplifted central part of a fold is the most developed, whereas the periclines lag behind. As a result an overthrust complicating the limb of a fold fades out towards the periclines. In this way, by comparing the cross-sections of a fold in the area of a pericline with cross-sections nearer to the middle it is possible to observe all the stages of transition of a bend into

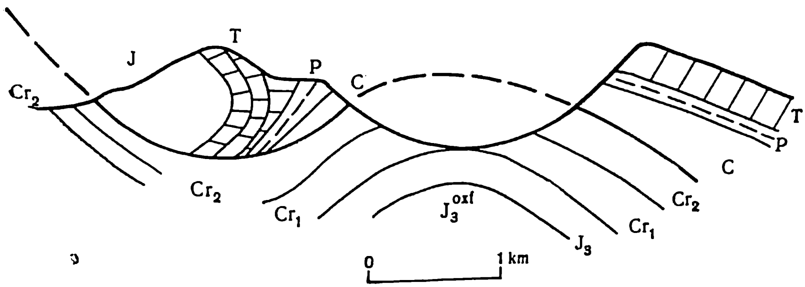


Fig. 93. Undulation of an overthrust caused by later folding (French Alps)

a fracture. It should be noted that the development of a thrust most often is preceded by the overturning of the fold: the overturned limb being the favourite place for the development of overthrusts. Furthermore, it is possible to observe gradual flattening of the underlimb and the extension of layers along the dip. This plastic extension, accompanied by boudinage, results in faulting in the form of an overthrust.

Since thrusts represent viscous fractures they tend to fade out along the strike. Furthermore, along the dip they split up into an increasing number of faults of diminishing throw. The surface of a thrust often shows waveform bending both in the horizontal and vertical directions. The undulation of an overthrust may be caused by a variety of factors. For example, the inclination of a thrust may vary as it passes from rocks possessing certain mechanical properties into weaker or stronger rocks. In plastic rocks a thrust is usually more gentle than in series of hard rocks. Undulation may also be due to the fact that an overthrust, like any dislocation, does not develop all at once, but, as it were, is "built up" out of small "incipient" fractures which develop separately and gradually merge together. Lastly, undulation may be secondary, that is, caused by later folding which occurred after the development of the overthrust (Fig. 93).

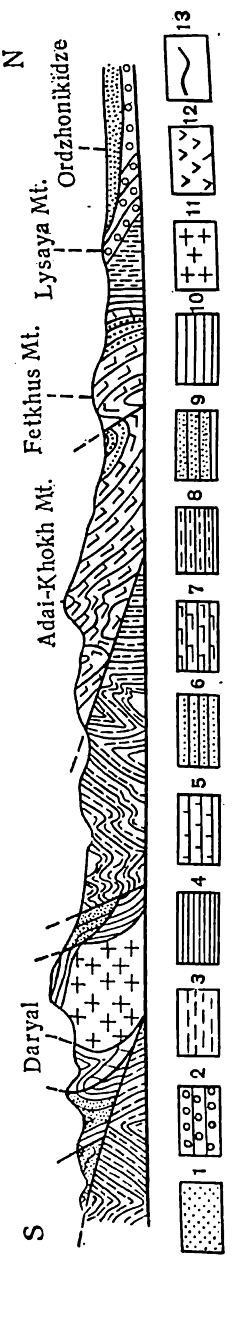


Fig. 94. Imbricate overthrusts of Central Caucasus (after V. Rengarten):
1—Akchagyl and Apsheron; *2*—Meothysian and Pontian; *3*—Upper and Middle Miocene; *4*—Paleogene; *5*—Upper Cretaceous; *6*—Lower Cretaceous; *7*—Upper Jurassic; *8*—Upper Lias; *9*—Middle Lias; *10*—Lower Lias; *11*—granites; *12*—dacites, andesites; *13*—faults

In regions where the folds have the same inclination the development of numerous overthrusts on the lower limbs of folds produces the so-called *imbricate structure* (Fig. 94).

The strike of overthrusts nearly coincides with the strike of the folding, deviating from it only in certain areas in connection with the general undulation of the overthrust surface. The inclination of the thrust in most cases closely approaches that of the axial plane of the fold. This, however, does not apply to the fan-like thrusts mentioned earlier.

Not all thrusts, however, are associated with specific folds. In folded regions there are often observed so-called *cross overthrusts* in which whole series of folds are thrust over another series. In such cross thrusts the displacement is generally in the direction of the regional dip of the folds, and the net displacement may reach several kilometres.

A special case are very large tectonic displacements called nappes or overthrust sheets, the morphology of which was discussed earlier.

At the beginning of this century, overthrust sheets were the vogue among geologists. Many of them believed that overthrust sheets were the most typical structure of mountain regions. It was assumed that overthrust sheets may have been displaced horizontally as much as two hundred kilometres. They contended that the greatest sheets were characteristic of the Alps, whose structure was conceived as a series of superposed sheets thrust over great distances. Furthermore, they maintained that overthrust sheets had formed out of recumbent folds whose underlying limbs are gradually attrited, and in some cases are destroyed altogether. The classic pattern of the successive stages of the development of over-

thrust sheets in conformity with this theory is given in Fig. 95.

The present views on the role of overthrust sheets in the structure of mountain regions is entirely different. More detailed and impartial studies have demonstrated that the importance of overthrust sheets was greatly exaggerated. Overthrust sheets associated with recumbent folds do exist in nature, but their net displacement nowhere

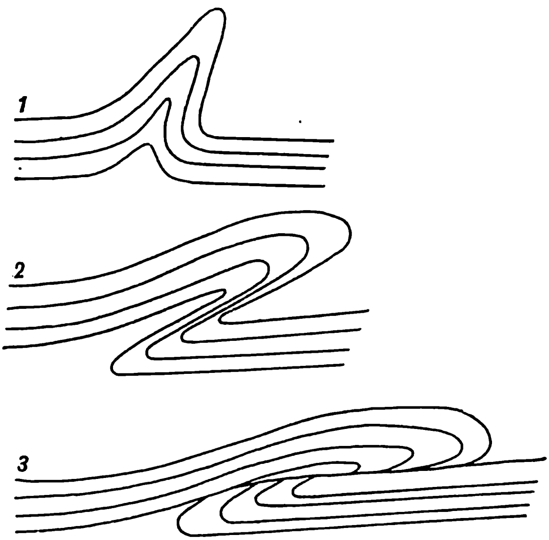


Fig. 95. Classic conception of development of an overthrust sheet from a recumbent fold:
1, 2, 3—successive stages of development of overthrust sheet

exceeds 10 or 15 km. An example of such an overthrust was given in Fig. 75. Horizontal displacements of greater amplitude are also known, but in the light of new data they are believed to be associated not with recumbent folds, but with the gliding of separate blocks of the earth's crust downslope of a mountain range under the action of gravity. Such a gravitational overthrust is shown in Fig. 96. In this case gliding was facilitated by the presence of a gypsiferous Triassic series composed of extremely "slippery" rocks at the foot of the block that has slid down. In the Western Alps, plastic Triassic strata and almost equally plastic Eocene flysch series generally play the dominant role in the development of gravitational overthrusts.

Occasionally the net slip of such overthrusts is as great as 40 km. Presumably such displacement did not occur all at once but in

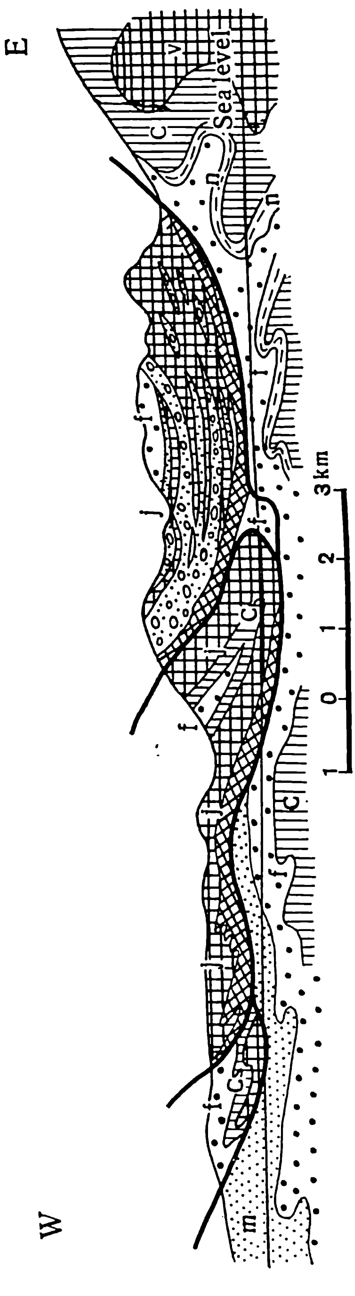


Fig. 96. Gravity overthrust in the forezone of the Savoy Alps: m—molasse; f—flysch; n—nummulite beds; C—Cretaceous; j—Jurassic; v—Triassic

stages as the range grew higher and wider and the tilted zone shifted towards the outward part of the range. Such sheets are completely detached from their "roots", that is the rock complex that remained in place. There are no reasons to believe that at any time after the development of the sheet its roots were directly connected with its body. The mechanism of development of such sheets is quite similar to the mechanism of development of landslides: some blocks of the earth's crust are detached and freely slide down by gravity.

The term "outlier of an overthrust sheet" mentioned above does not apply to sheets of this type, inasmuch as even originally they constitute a complex of blocks that have detached from their roots (and from each other) and slipped downslope.

The possibility of the development of such big landslides on gently-dipping slopes generally characterizing mountain ranges is open to doubt. However, in part they may be explained by the earlier-mentioned phenomenon of creep. Plastic rocks that lie on slopes for great lengths of time very slowly flow down. It seems that such creep could develop on slopes which do not exceed a few degrees.

Gravitational overthrusts constitute a phenomenon in which the properties of both surficial (non-tectonic) and tectonic processes are combined. A gravity overthrust resembles a landslide but it is of such a scale and

accompanied by such additional phenomena (crumpling of strata into major folds, development of friction breccia, etc.), that it would be strange indeed to regard it otherwise than a manifestation of the tectonic life of the earth's crust.

The sole of the sheet is often undulating. Sometimes this undulation is primary, that is due to the fact that the sheet had moved over an irregular surface, but more frequently is the result of later dislocations (development of dome-shaped uplifts and basin-like depressions) that developed regardless of the presence of the overthrust sheet and had entrained it and deformed its sole.

The strata which form the sheet are often crumpled into numerous irregular folds and broken by faults of various types.

Compression of rocks also results in strike-slip faults. In an overthrust the given segment of the earth's crust contracts horizontally (across the strike of the folds) and grows thicker vertically: a shift characterized by horizontal slip parallel to a vertical or a steep fracture, also is a shear. In this case, however, the given segment of the earth's crust both contracts and grows thicker along the strike of the fold.

Thus, while overthrust sheets strike in a direction closely coinciding with the strike of folds, strike-slip faults that involve crumpling, strike at an oblique angle to the folds. Displacement along overthrusts and strike-slip faults continues (as far as the direction of the principal axes is concerned) the deformation begun by folding (Fig. 97).

Right-hand and left-hand strike-slip faults are recognized. In a right-hand strike-slip fault the wall farthest away from the observer is displaced to the right. In a left-hand strike-slip fault the picture is reversed. It will be easily seen from Fig. 97, that the designation of a strike-slip fault as a right-hand or left-hand one does not depend on the direction it is viewed from.

Discussing in this chapter the relationship between folds and faults we proceeded from the assumption that thrusts and strike-slip faults develop after folding and constitute its continuation. Instances are known, however, when faults of the steep overthrust type clearly anteceded folds. Apparently this was due to such mechanical properties of the rocks and such character of the forces that produced conditions which were more favourable to faulting than to folding. Faulting, however, weakens the rock series and produces slip planes within it. This favours the development of plastic deformation in the form of folds, which in such cases develop near the faults. In these folds the strata are bent as a result of friction developing when the rocks slip along the fault planes.

Faulting associated with shear deformation of the earth's crust may take the form of normal and reverse faults and strike-slip faults. Normal and reverse faults are produced when the couple of forces

that caused the shear deformation was oriented vertically and resulted in relative uplifting and subsidence of segments of the earth's crust.

If a segment of the earth's crust rises along a vertical fracture relatively to adjacent segments, a horst bounded by faults is formed. The fissures may be not vertical but inclined, with the fault

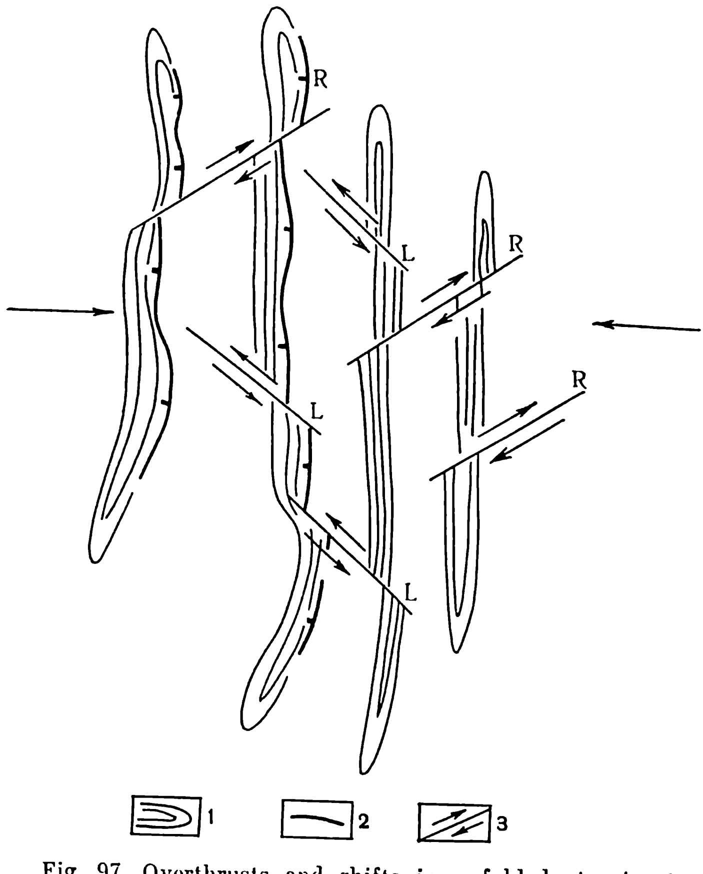


Fig. 97. Overthrusts and shifts in a folded structure:
1—contour lines of anticlinal folds; *2*—overthrusts; *3*—strike-slip fault (*R*—right-hand strike-slip faults; *L*—left-hand strike-slip faults)

plane dipping towards the foot wall (normal faults) but often they have a reverse dip beneath the hanging wall in which case they resemble reverse faults. Very common are horsts of fan-like structure, bounded on either side by faults dipping towards each other. Such are, for example, the numerous horsts of the Tien Shan (Fig. 98). There are reasons to believe that such dipping of the fissures bounding

the horsts is secondary and is associated with the spreading of the upper part of the upthrust block mainly under the action of gravity. Cases are known when the fault fissures are continually deflected, their dip becoming more and more gentle upward so that eventually they become horizontal. Such upward flattening out of fissures in the course of which a fault passes into a low-dipping overthrust may cause the splitting of the upthrust mass and caving of some of its blocks. However, a vertical position should be inferred for the deeper zones of such fissures (Fig. 99).

Naturally, the fault planes at the sides of a horst may be asymmetrical (vertical on one side and inclined on the other). Other combinations are possible as well. Horsts associated with an uplift may be unilateral similarly to unilateral box-folds (see p. 61). In other cases they may be complicated by step faults of the second order.

Uplifting of certain segments of the earth's crust along faults goes parallel with subsidence of other segments. Often the upthrust and downthrust blocks alternate like keys.

In keeping with the above, the upper parts of faults bounding downthrust blocks tend to dip away from the downthrust trough towards the adjacent masses. The latter, spreading out under the action of gravity, tend to fill the trough and creep on it.

The couple of forces that causes shear in the earth's crust may be directed not vertically but horizontally or obliquely and this is how strike-slip faults are produced.

For example, when complete folds are formed, sometimes one part of the fold is displaced relatively to another in the horizontal direction along strike-slip faults which run transversely to the strike of the folds. In most cases this displacement is not a pure strike-slip fault but one with

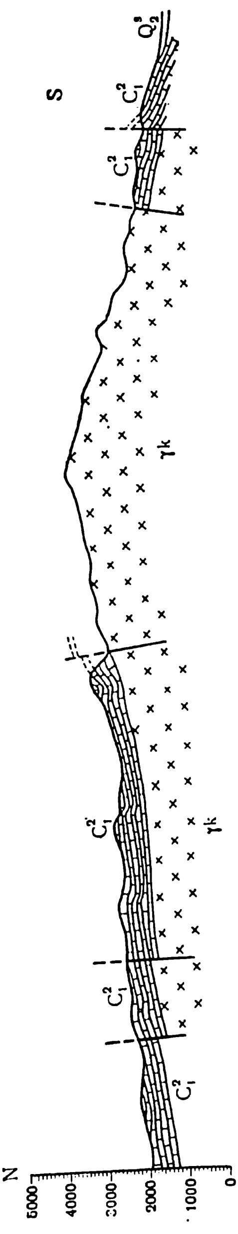


Fig. 98. A horst with steep fan-like overthrusting in the Trans-Ili Alatau of Tien Shan (after A. Goryachev)

some vertical throw. The rocks involved in a thrust fault or sheet overthrust may be divided into blocks by normal faults that run across the strike, the individual blocks being displaced along these faults in a horizontal or nearly-horizontal direction. As a result, some parts of the overthrust mass may be pushed farther ahead while others may lag behind.

Peculiar forms of overthrusting are observed in regions where the strata are crumpled into folds with vertical or very steep axes.

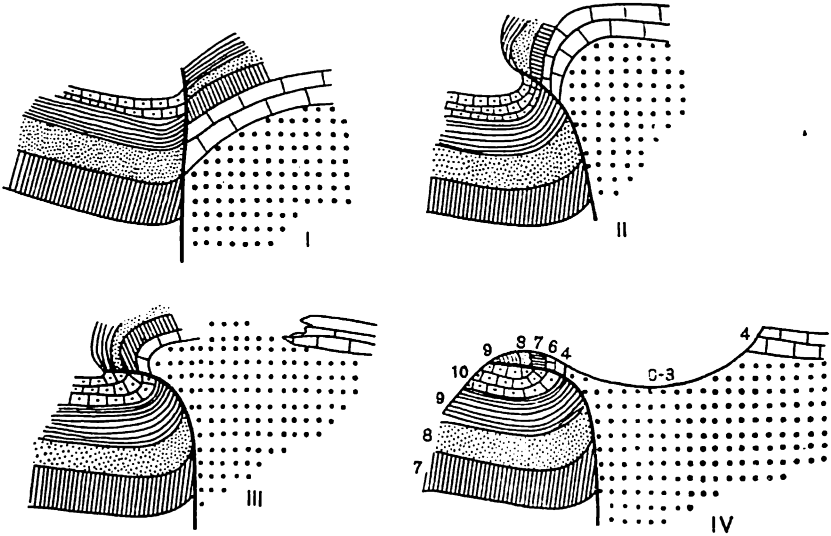


Fig. 99. A fault passing at the top into an overthrust. Sections of the same dislocation at different stages of bending (Jura Mts.): dotted—Triassic; figures from 4 on—Jurassic horizons from the bottom up (after Glangeout)

Such conditions are very common in Archean series on ancient shields. Figure 100 shows the minor folding in the Archean gneiss series of Karelia. Unless one notices that the drawing represents a horizontal section the natural assumption would be that the faults complicating these folds are reverse faults. Considering, however, that the axes of the folds are vertical, according to the terms accepted in geology, these faults should be regarded as strike-slip faults. The mechanism of their development in this case evidently does not differ from that of overthrusts associated with folding.

A special problem is that of existence in the earth's crust of major horizontal displacements which are important on their own. Some (mostly American) researchers admit the possibility of existence in

the earth's crust of horizontal displacements with a net amplitude of hundreds and even more kilometres and assign to them a very important part in the tectonic life of the crust. Conversely, others maintain that horizontal displacements are always subordinate to vertical ones. This problem is not yet, however, completely solved.

In 1906 a catastrophic earthquake occurred in the area of San Francisco and as a result the greater part of the city was destroyed. It has been established that the earthquake involved a horizontal displacement with a maximum throw of 7 m along a vertical fracture of the great *San Andreas fault*. Subsequently, some geologists found

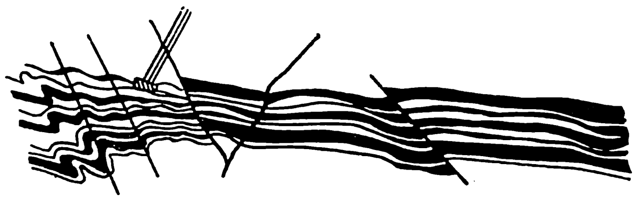


Fig. 100. Faults in folds with vertical axial planes in the Archean rocks of Karelia (after A. Sorsky)

certain structural evidence near this fault showing that such horizontal displacements had occurred here earlier, during older geological periods, and that their total slip amounted to hundreds of kilometres. In this way strike-slip faults became the vogue, as overthrust sheets had been a few decades earlier. Certain geologists found evidence of similar movements along other faults as well. The Pacific coast was especially "lucky" in this respect. Some authors contended, for example, that the entire floor of the Pacific Ocean together with the adjacent littoral parts of North America, the Japanese Islands, New Zealand, etc. for several geological periods had been rotating counter-clockwise relatively to the inner parts of the adjacent continents.

Other researchers, however, maintain that the movements along faults encircling the Pacific Ocean were mainly vertical and only occasionally involved purely subordinate horizontal displacements.

This problem is basic to geotectonics, because on its solution depends the understanding of the processes taking place in the deep interior of the earth, the factors that cause tectonic movements, etc.

All shear deformations are accompanied by development of additional fracturing, such as echeloned fractures both of the tension and shear types mentioned earlier.

22. KINEMATIC CLASSIFICATION OF FAULTS

The above classification of faults was based on morphological features. However, considering the above data on the conditions of development of faulting, faults may also be classified according to their association with certain kinds of crustal movements. Such a classification would be essentially a kinematic one and thus constitute the first step towards a genetic classification of faults.

The following types of movements are recognized: tension, compression, uplifting and subsidence. The last two types of movement are open to doubt since uplifting and subsidence are always relative and in each given case it is difficult to tell whether section *A* has been uplifted relatively to section *B* or vice versa. Such difficulties do arise but nevertheless in most cases this problem can be positively solved. For example, if amidst a vast region of horizontal strata rises alone tectonic brachycline this will plainly be a case of uplifting of the brachycline not of subsidence of the entire vast plain relatively to it. Similar cases could be mentioned with regard to subsidences as well.

Taking this approach to the problem, it may be suggested that all faults associated with tension in the earth's crust and subsidence of its separate blocks be described as normal faults. In this case reverse faults will be those that develop in the course of uplifting of a segment of the earth's crust. Overthrusts and strike-slip faults will then be associated with compression.

In this classification of faults, the inclination of the fault plane is not decisive. In the morphological classification, for example, the difference between reverse and normal faults is that in the case of the former the hanging wall is higher than the foot wall. But it is conceivable that the uplifting of a tract of the earth's crust may result in a fault whose foot wall is higher. Under the morphological classification this is a normal fault, while under the kinematic classification it is a reverse fault.

The same applies to the relationship between a reverse fault and an overthrust. It may be said that overthrusts are always associated with compression primarily expressed in folding. Therefore, if it can be established that a given fault has formed under the conditions of horizontal compression and is associated with displacement along the dip of the fault plane, it should be referred to as an overthrust, even if the fault plane dips at an angle of more than 60°.

The kinematic classification of faults is not complete, and we mentioned it here to outline the possible ways for developing it further. For the time being, in practical work it is more convenient to use the morphological classification.

23. DEEP FRACTURES AND THE FUNDAMENTAL DIVISIBILITY OF THE EARTH'S CRUST

Soviet geologists have introduced into science the concept of so-called "deep fractures" of the earth's crust, that is tectonic fractures of great length and depth which probably go through the entire crust. They have a very long history which means that displacements along deep fractures apparently were always vertical, occurred over great lengths of geologic time and were renewed many times. Furthermore, the direction of the displacement was often reversed. Up and down movements of segments of the earth's crust along such fractures could have occurred through whole geological periods and even eras. Thus, the movements along deep fractures constitute a manifestation of the continuous oscillatory movements of the earth's crust. In this case, however, the regions of uplifting and subsidence are linked with each other not by smooth flexures but by fractures in the crust. As in other cases of up and down crustal movements, the segments separated by deep fractures, depending on the nature of their movements (rate, throw, etc.) differ between themselves in stratigraphic sections, thickness of strata of the same age, their facies, stratigraphic intervals, etc. It has been mentioned earlier that all sedimentary series change areally both with regard to thickness and facies. However, this applies to gradual changes. In the case in question the changes in thickness and facies along the deep fractures are abrupt. In many cases the geological sections opposing each other along such fractures radically differ between themselves. Formerly such close proximity of sedimentary series of the same age but belonging to different facies and distinguished by different thicknesses was explained by the missing of transitional sections, which were supposed to have been buried in the course of development of overthrust sheets. These sheets in turn were regarded as a result of contraction of the earth's surface. In conformity with this interpretation, many deep fractures were regarded as tectonic thrust surfaces of a great scale. As to the position of the fractures, it was inferred that it was vertical only at the surface of the earth and that at a certain depth they became low-dipping or even horizontal. Now it has been positively established that in most cases these views are erroneous.

Morphologically, deep-seated fractures constitute major vertical faults, accompanied by crush zones and tectonic breccias.

The problem of deep-seated fractures is directly related to that of the "fundamental divisibility" of the earth's crust. Figure 101 is an aerial photograph of a part of the Baltic crystalline shield. It will be seen that here the earth's crust is broken by numerous fractures trending in various directions. It is not possible, however,

to relate this fracturing with crustal movement of a more general nature, for example, with the shield-like uplifting of Scandinavia (the Baltic Shield). Such division of the earth's crust exists elsewhere too, and there are reasons to believe that it is universal, though in different regions it is recorded in different forms. It is most clearly in evidence where the surface of the earth is composed of crystalline rocks, but is observed in dense sedimentary rocks as well. Certain regular dependences are usually observed in the location of such fractures. For example, in the USA, on a vast area of the Colorado

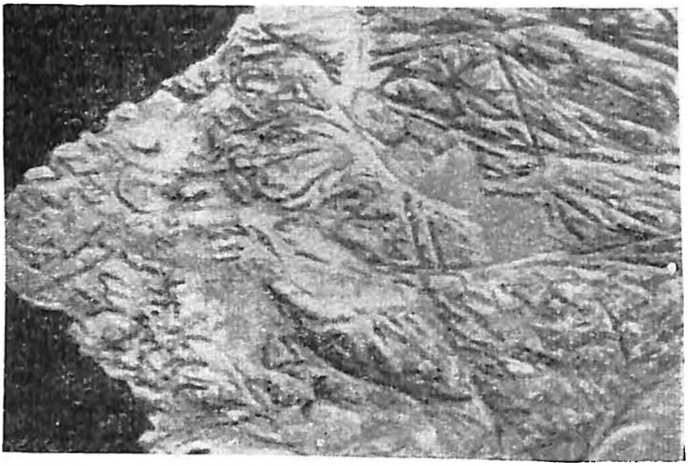


Fig. 101. Aerial photograph of a part of the Baltic Shield showing regional fractures

Plateau, rocks of most diverse mode of occurrence are traversed by fractures mostly trending in a meridional direction. Most of fractures of the Russian Platform have a northeastern or a northwestern trend.

Often the "fundamental divisibility" of the earth's crust is recorded in drainage. In some regions of the Hungarian Lowland, which as known occupies an ancient mass composed of metamorphosed and intensely dislocated Paleozoic and older rocks mantled with a thin cover of loose sediments, an amazing regularity is observed in the location of rivers, which flow mostly in a meridional direction, though some of their sections are latitudinal. Presumably this river pattern reflects an orthogonal system of regional fractures characterizing this area. The regional fractures in the folded basement are also reflected in the sedimentary mantle, producing weakened directions, which are later exploited by river erosion.

It was suggested that on platforms the general fractures the origin of which is uncertain constitute a reflection of the same regional fractures in the sedimentary cover which explains their regular pattern. However the mechanism by which regional fractures are "reflected" in intrastratal general fractures still remains an unsettled problem.

Another noteworthy fact is that in some regions of the globe major tectonic elements exhibit a regular strike. For example, the entire great belt that comprises the Eastern Sayany, the Altai, Central Kazakhstan, Tien Shan, Kopet-Dagh, the Caucasus, a part of Europe north of the Carpathians and the Alps is plainly characterized by a northwestern strike of the zones of lifting and subsidence of the earth's crust over great lengths of geological time. Furthermore, it involves both geosynclinal and platform tectonic elements.

This warrants the suggestion that regional divisibility of the earth's crust is universal and is exploited in all tectonic crustal movements, either local or those that involve vast areas, i.e., the regional fractures determine the boundaries of tectonic zones, of regions of uplifting and.subsidence, etc.

The problem of deep fractures of the earth's crust is one of the basic p?oblems of modern geotectonics which requires much further study. Until now there is no settled opinion concerning the existence and the strike of such fractures and related regional fractures involving whole continents. The origin of regional fractures is the subject of various vague conjectures.

However, the very fact of existence of regional division of the earth's crust is no longer questioned and must be taken into account to explain certain specific disjunctive dislocations. For example, earlier we have mentioned the regular pattern of fractures on the arches of tectonic uplifts (domes, brachyclines, anteclises). In these cases the pattern is entirely controlled by the form of the uplifts. It was noted that the mechanical non-uniformity of the rocks that form the uplift may distort the pattern of fractures. Now we should add to this that in some cases the position of faults (usually of the normal type) on uplifts is not controlled at all by the shape of the uplift but by regional fracturing, which is exploited and rejuvenated in the course of the growth of a given uplift.

Such an uplift may be broken into a bizzare mosaic of blocks, whose pattern does not depend on the shape of the uplift.

24. PHENOMENA ACCOMPANYING FAULTS

Faulting is accompanied by a number of phenomena reflected in various alterations in the structure of the rocks adjoining the fault.

Usually only the smallest tectonic breaks in continuity are expressed by a single fracture. Even then it splits at the ends into thin

cracks. Major tectonic breaks are always represented by a rupture zone whose width may vary from fractions of a metre to hundreds of metres. This zone comprises numerous fractures, which for the most part are roughly parallel but may also cross each other at acute angles. As a result the rupture zone is usually broken up into narrow long lenses of relatively undisturbed rocks fringed by intensely crushed bands. The width of the zone of fracturing and attrition along a fault varies widely, sometimes to the point of its complete disappearance.

The displacement of rocks along the fault plane is accompanied by friction, which produces scratches, furrows and gashes along the walls of the fault. These marks indicate direction of the displacement. Often mirror-smooth surfaces, called *slickenslides*, develop along the fault plane as a result of the polishing action of the sliding rocks and the formation of new flat brilliant minerals in the friction zone (chlorite, sericite, epidote, etc.).

Friction also causes crushing and attrition of rocks along the fault plane. Crushing primarily causes intense fracturing, in view of which the zone adjoining the fault is always characterized by more numerous fractures and systems thereof.

Another effect of faulting is the so-called *tectonic breccia* which develops as a result of friction when a block criss-crossed by numerous fissures is displaced and turned about. The blocks in tectonic breccias show most diverse forms, but most often are rounded and oriented parallel to the fault plane. The blocks in tectonic breccias sometimes measure tens of metres across but usually range from a few centimetres to a few dozens of centimetres. Slickenslides and grooves are observed on the surface of the blocks. In some cases tectonic breccia resembles a coarse glacial moraine.

Fine tectonic breccia, called *kakirite*, consists of fragments less than one centimetre across. A big lump of kakirite crumbles in the hand into fine angular or lenticular fragments with gliding surfaces.

Still finer breccias composed of microscopic fragments are referred to as *cataclasites* and the process of their development as *cataclasis*. Under the microscope cataclasites appear as large crushed quartz grains surrounded by a halo of crushed quartz "fines". Other minerals too are crushed and bent.

Further crushing and attrition produces *mylonites*—rocks whose grains have been crushed to dustlike dimensions. Under very high magnification it can be seen that these grains are very flat and lenticular. Outwardly mylonites constitute a dense hornfels-like rock with a ribbon or fibrous texture.

The development of various tectonites along the fault plane is controlled by several factors, such as net slip, the depth at which the displacement occurred, the mechanical properties of the rocks,

and the rate of deformation. Depending on different combinations of these factors along a fault, the types of tectonites and the width of the zone occupied by them vary widely from place to place. However, special features distinguish tectonites associated with tension ruptures from those typical of compression ruptures. Tension faults are mostly associated with broad bands of coarse angular breccias. Thrust faults, on the other hand, are usually characterized by relatively thin zones of finely attrited rocks—cataclasites and mylonites; furthermore, thrust fault breccias are usually distinguished by elongated, oriented and "compressed" blocks.

25. STUDY OF FAULTS

Major faults are recorded in the course of geologic surveys, geologic mapping and profiling conducted on different scales. What has been said above about the study of folded structures fully applies to disjunctive dislocations. Correct conclusions on the nature and mechanism of their formation require detailed investigations to record all their structural features. In such investigations it is essential to adhere strictly to relative scales for recording phenomena of different orders. This includes investigation of the fault zone from the point of view of the individual faults of which it is composed, the development of tectonites of various kinds and minor fracturing. Often the direction of displacement along the fault is indicated by the grooves along slickenslides. This method, however, should be used with caution. For example, a major fault whose great vertical displacement is established on the basis of general geological data, develops by a series of separate movements, whose directions need not have been vertical in every case. Certain displacements may have been inclined or even horizontal and the vertical displacement may represent their net result. Therefore, the orientation of scars and grooves which are always related to some specific movement may be at variance with the direction of the net displacement. Furthermore, in major fault zones displacement generally takes place along a number of partial faults simultaneously and such partial displacements may develop in different directions. Some huge blocks in the fault zone are often turned around. Therefore the grooves, even on an individual block, may run in different directions.

Thus, the direction of displacement can be positively inferred only by using the general geological criterion based on the relationship between the rocks on either side of the fault and on the net throw of the fault which is established geologically.

Inasmuch as faulting is closely associated with other kinds of dislocations (folds of different types, anteclises, etc.) faults cannot be studied out of the context of the entire tectonic environment.

In every case the relationship between faults and other tectonic dislocations should be ascertained.

It was noted above that the statistical method, based on a large number of measurements of the orientation of joints is widely employed in their investigation. Generalization of measurements gives a good idea of the predominant trend of the joints. When joints in a stratified sedimentary series are studied it is essential to determine then orientation relatively to the strata. In this way it is possible to identify the set of joints that is not related to the bedding of strata and cut across them regardless of folding, and also the set of joints whose position is related to the attitude of the strata and changes with the inclination of a given stratum. A point to remember is that statistical methods should be applied to joints occurring in the same geological setting, for example, within a single limb of a fold, or a pericline. If data on joints occurring in different geological settings are processed together, the resulting picture may be confusing.

The orientation of joints is very important for reconstructing the tectonic stress fields. Tension joints determine unmistakably the direction of the principal axis of tension since they are perpendicular to it. However, on the basis of tension joints alone it is not possible to reconstruct the principal direction of compression, inasmuch as it may lie in any direction parallel to these joints. The plane of the principal stresses is determined by shear joints. This plane is perpendicular to the line of intersection of conjugated shear joints. In this way tension and shear joints together completely determine the tectonic stress field. Nevertheless, this field can be reconstructed from shear joints alone by taking advantage of displacements along them, however insignificant, and also of the deviation of conjugated joints from the right angle.

The direction of displacements along shear joints completely determines the position of the axes of maximum compression and extension (Fig. 102). Furthermore, as we have said earlier, the orientation of shear joints as a rule deviates from the theoretical one. Therefore, conjugated joints may meet not at right angles but at blunt or acute angles. The bisectrix of blunt angles corresponds to the axis of maximum tension and that of acute angles to the axis of maximum compression. These considerations do not allow, however, for the possible effects of the nonhomogeneous environment on the orientation of the joints.

Two systems of conjugated shear joints and one system of tension joints may develop in a given stress field, inasmuch as the first are associated with maximum tangential stresses and the second with maximum normal stresses. In the absence of certain special conditions the development in the same stress field of joints of different

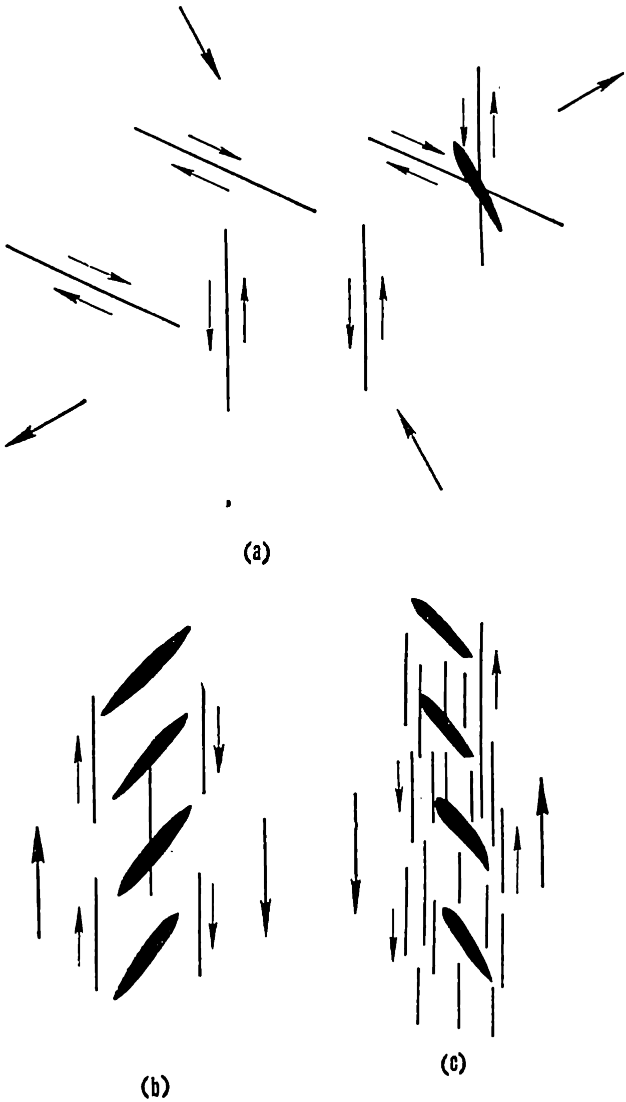


Fig. 102. Reconstruction of tectonic stress fields from fractures: *a*—in the case of compression; *b*—in the case of right-hand and *c*—in the case of left-hand strike-slip faults. Thick lines—tension joints; thin lines with arrows—shear joints

directions corresponding to some intermediate values of tangential and normal stresses is a mechanical impossibility. Such conditions may be stress concentrations around openings in rocks, natural or artificial, or local nonhomogeneities in the environment.

Lacking such conditions, the existence of several systems of conjugated shear joints indicates that these joints developed at different times and that the stress field changed with time as well.

Reorientation of the stress field may result in a change in the nature of the joints. For example, under new conditions shear joints may pass into tension joints and open up, whilst tension joints in the new stress field may play the role of shear joints.

Inasmuch as stresses cause not only jointing but plastic deformation as well, the reconstruction of the stress field from joints may be supplemented by observations of plastic deformations. For example, folds of general crumpling develop at right angles to the axis of maximum compression. As we have said earlier, the direction of compression can be reconstructed from the deformations of fossil organisms, oolitic and other grains, cleavage, etc.

In every case of reconstruction of tectonic stress fields one must bear in mind that as soon as certain stresses cause plastic deformation or ruptures, the stress field changes. Therefore, in the deformed medium the stress field may vary from place to place in a complicated manner. It can be easily demonstrated that in different parts of a developing fold the directions of stress axes are entirely different. In view of this stress fields must be studied locally, from one area to another, which means that it is not possible to ascertain the stress field of a big area only on the basis of observations made at individual points.

CHAPTER VII

Dislocation of Igneous and Metamorphic Rocks

26. DISLOCATIONS OF IGNEOUS ROCKS

Our previous discussion of various tectonic dislocations was mostly concerned with stratified sedimentary rocks. Now we shall make certain points regarding manifestations of dislocations in igneous rock masses.

Under the action of tectonic forces effusive rocks occurring as beds between sedimentary rocks and also sheets of intrusive rocks suffer deformation approximately in the same way as stratified sedimentary series. Generally the dislocations developing in such bodies do not differ in principle from dislocations in sedimentary series, though there are certain specific features that depend on the different mechanical properties of sedimentary and igneous rocks.

In a zone of intense folding, sheet-like igneous bodies most often crack and break up. Therefore, sometimes instead of solid sheet-like masses separate lenses of igneous material are observed.

The picture is more complicated in the case of tectonic deformations of massive intrusive rocks.

Some geologists contend that dislocation of intrusive masses, especially large ones without a break in continuity is impossible and they can only split under the action of tectonic pressure. This is not so. It is known from observations that, given appropriate conditions, intrusive rocks can suffer intense plastic deformation without a break in continuity.

It should be borne in mind, however, that practically always intrusive rocks are less plastic than sedimentary rocks, which determines their behaviour relatively to the surrounding sedimentary rocks in the process of deformation. Therefore, when a sedimentary series undergoes plastic deformation, intrusive masses most often split. However, in the conditions of high temperatures, confining pressure, and a slow deformation rate an intrusive mass can "flow" almost in the same way as the enclosing sedimentary rocks.

Furthermore, the result of deformation depends both on the initial shape of the body suffering deformation and its internal struc-

ture. Under longitudinal compression a bedded sedimentary series is crumpled into folds, but intrusive rocks which initially were massive and had a homogeneous internal structure naturally could not develop such secondary forms. Furthermore, deformations which are plainly observable in the case of folded sedimentary rocks, are much more hard to detect in massive rocks, whose original mode of occurrence cannot be easily reconstructed.

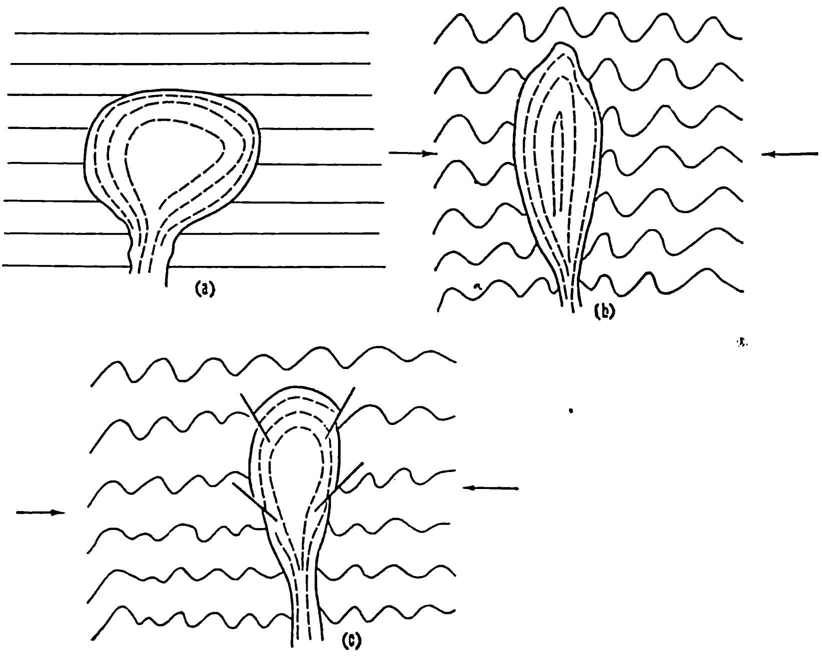


Fig. 103. Deformation of an intrusive mass in bedded sedimentary rocks: *a*—intrusion in undisturbed strata; *b*—the same intrusion subjected to compression in the course of folding of enclosing bedded rocks; *c*—intrusion that suffered fracturing under the same conditions

Let us consider an intrusive mass occurring in bedded rocks (Fig. 103*a*). When the earth's crust in the given area is subjected to horizontal compression the sedimentary rocks are crumpled into folds and the homogeneous intrusive mass is deformed as a single body, being compressed horizontally and extended upwards accordingly (Fig. 103*b*). This goes along with mechanical detachment of the intrusive mass from the enclosing rocks expressed in shearing along the contact. Such shearing between sedimentary and massive intrusive rocks is regularly observed in regions of dislocations.

The shape of the mass is not what it was originally, but this is very hard to prove. The same figure (Fig. 103*c*) shows an intrusive

mass which under the conditions of folding of the enclosing rocks proved to be incapable of plastic deformation and split. Such dislocation, of course, can be more easily observed and analyzed.

If, however, a disjunctive dislocation occurs inside the intrusive mass it is rather difficult to determine its throw and direction. To do that it is necessary to locate the place where the given break displaces the outer contour of the mass and extends into the enclosing rocks.

Studies of deformations of intrusive masses are facilitated by investigation of their internal structure.

It is known that in the process of tectonic deformation of intrusive masses their internal structure suffers certain changes. The combined effects of pressure and movements occurring in the rock are expressed in changes in the shape and orientation of the grains and partly in their composition. The entire complex of these phenomena is known as *dynamic metamorphism*.

The secondary internal structure of igneous bodies expressed in linear, plane-parallel or linear-parallel orientation of the crystals generally has the same character as the primary structure. Therefore, at the first glance it is easy to confuse these structures. Such confusion can be avoided by deeper investigation of the internal structure. The prime requisite is a clear understanding of the mechanism by which tectonic forces affect the arrangement of crystals within a rock.

Changes in the arrangement of crystals may be caused by various factors. First, flat or elongated crystals may be turned and become parallel to each other. This happens when deformation involves relative sliding of separate parallel slabs. However, from the entire preceding discussion it is clear that such sliding develops in any rock suffering plastic deformation. Furthermore, this sliding develops along the planes of maximum tangential stresses. For example if a sheet-like igneous body is bent into a fold, due to sliding of the layers relatively to one another, in addition to being bent such a body is subjected to shear deformation. As a result an internal plane-parallel structure lying in the plane of the entire sheet-like body may develop. Conditions most favourable to the development of such a structure will be observed at the boundaries of the body, that is at its foot and roof.

Second, the orientation of crystals may be due to their deformation by the same mechanism that was described for flow cleavage. In this case the plane-parallel orientation of the crystals is normal to the direction of maximum compression. If an irregular igneous body is subjected to horizontal compression this mechanism will produce a plane-parallel orientation that will be vertical and parallel to the strike of the folding in the enclosing bedded rocks. If

the folds are inclined, such an orientation, associated with deformation of crystals, is generally parallel to the axial planes of the folds.

Microscopic examination shows that on compressive deformation many crystals are crushed and later recrystallize in a new direction.

Third, crystals are oriented as a result of unilateral growth in a stress field. In keeping with the so-called Riecke principle, if a crystal, which is in contact with a solvent, is subjected to compression in a certain direction, the crystal substance on the faces under compression is dissolved and transported to the faces that do not experience compression and is deposited there. In this way the crystal grows shorter along the axis of compression and longer along the axis of tension, which produces a corresponding orientation of the crystal.

Considering that the two last factors have the same end result, we are entitled to recognize only the two principal mechanisms of development of secondary oriented structures: by rotation when crystals turn parallel to the planes of maximum shear stresses; and by deformation and recrystallization, when the crystals are oriented at right angles to the axis of maximum compression and parallel to the axis of maximum tension.

In every case a secondary oriented structure is distinguished from a primary structure by the fact that outwardly it shows a most close relationship to the deformations in the enclosing rocks, inasmuch as it is produced in the process of deformation in which both the igneous mass and the enclosing rocks are involved. For example, a plane-parallel structure produced by deformation or recrystallization will be parallel to flow cleavage in the enclosing folded bedded rocks. Conversely, the primary internal structure of intrusive masses largely depends on the mechanism of formation of the mass itself, specifically on the process of movement of the fluid or semi-fluid magma that fills the given space. Therefore, the nature of a given internal structure can be established best on the basis of a combined study of the internal structure of the intrusive mass and of the tectonic structure of the enclosing rocks.

Approximately the same applies to fissures observed in intrusive masses. Previously we spoke of primary fissures in intrusive masses whose position is controlled by the internal structure, whether linear or plane-parallel. When such a mass is subjected to the action of tectonic forces, depending on the stress field, it develops either tension or shear fractures, which are closely related to the systems of similar fractures that have formed in the enclosing rocks during the same process of deformation. Fig. 104 shows an intrusive mass in the core of gentle arch uplift. The usual complex of primary fractures is observed within the mass, but in addition it also has fractures that are radial relatively to the uplift and cut across both the mass

and the enclosing rocks. These are typical secondary tension fractures produced by bending, that have developed in the course of general deformation of the intrusive mass and the enclosing rocks.

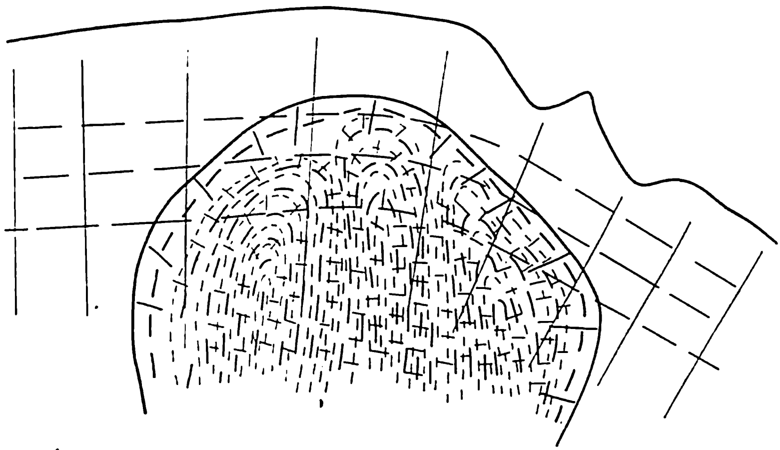


Fig. 104. An intrusive mass broken by internal and external fractures (common to the mass and to the enclosing rocks):
short thin lines in the mass represent oriented structure; thick lines—internal fractures; solid thin lines—external tension fractures, radial to the bending of the strata; long dashes—"surficial" cracks outlining the mass

We examined certain special features of the dislocation of igneous rocks by external tectonic forces under conditions when the part played by the rocks was passive.

Fig. 105. Formation of a caldera. Lava is shown black

In other instances, however, magma may be an active agent. Moving in the earth's crust it exerts a mechanical action on the enclosing rocks causing various deformations. In many cases the effect of magma on the enclosing rocks is similar to the effects of salt or gypsum diapir cores. On intruding laccoliths, igneous diapirs and

stocks lift the overlying strata, bend them into domes and push them apart. Often the strata uplifted and bent by the intruding magma are ruptured and then the dome-shaped uplifts are complicated by fractures which are controlled by the same dependences as fractures on diapir domes. In some instances the magma pierces the roof and reaches the surface to form bodies that are partly intrusive and partly effusive.

A class apart are dislocations associated with processes developing in the hearths of active volcanoes. In some instances eruption of large quantities of lava and gases results in the subsidence of large sections of the earth's crust. In this way are formed *calderas* or *cauldrons* which may be tens of kilometres in diameter. From the purely tectonic point of view such calderas are circular fault troughs bounded by faults (Fig. 105).

27. DISLOCATIONS OF METAMORPHIC ROCKS

Most metamorphic rocks have derived from the alteration of sedimentary rocks. In view of that they generally retain elements of stratification which often are so pronounced that they retain such discernible elements as rhythmic or cross-bedded lamination, ripple marks and other features indicating the conditions under which the original sedimentary rocks were deposited.

The bedding of metamorphic rocks is seldom undisturbed, although of late several instances of nearly horizontal occurrence of metamorphic rocks on vast areas have been revealed. As a rule, metamorphic rocks are strongly dislocated.

However, the dislocation of metamorphic series is somewhat specific. In typical cases metamorphic rocks exhibit great plasticity or, to be more exact, flowage in the process of deformation. The folds developing in metamorphic series in most cases are very complex and represent folding of many orders. The limbs of major folds are generally complicated by minor folding, which in turn is complicated by still smaller folds and so forth to the point when puckering, that is tiny folds measured in terms of centimetres, are produced. Furthermore, this is accompanied by sharply pronounced disharmony due to non-uniform flowage of the material of the beds and changes in their original thickness in the course of which material was squeezed from some places and injected into others. The impression, which is obviously true is that in the process of deformation the material of metamorphic rocks was in such a flowable state that it reacted to stresses by most non-uniform flowage. Furthermore, the great flowability of the material explains the almost complete absence of major faults. Small faults do exist but rapidly pass into plastic deformations in all directions. Whenever metamorphic series

are split by major tectonic faults, it can always be established that these have developed much later in entirely different tectonic conditions and are not related to the dislocations observed in the metamorphic series itself. On the Baltic and Canadian shields, for example, which are composed of Precambrian rocks, major regional faults are Tertiary and not related to Precambrian dislocations. Some of these faults seem to be associated with young uparching of these shields.

Very characteristic of metamorphic Archean series is the absence of major linear folds. These series in the main appear as great round, oval or irregularly shaped domes, which in some instances measure tens of kilometres across. Often at the core of such a dome lies a batholith or an intrusive gneiss body enveloped on the limbs by steeply-dipping metamorphic and crystalline schists. In most cases the limbs of big domes are complicated by radial folds of different orders, characterized by very steep, up to vertical, axial planes. Reversed axes forming pseudosynclines and pseudoanticlines are also observed here (Fig. 106).

It may be assumed that the great domes in the Archean series were produced by the direct action of granite intrusions which lifted the overlying rocks. The overlying strata were bent upward until their position became vertical, whereupon they found themselves under compression directed normal to the strata inasmuch as they were caught in the intervals between the rising intrusions. Naturally this compression was non-uniform, being stronger at some places and weaker at others. This caused intrastratal movement of material which was squeezed from some places and injected into others. This produced minor folds with steep axial planes as a result of local "piling up" of material and its "overrunning" in the process of intrastratal movement. The folding was accompanied by extensive boudinage, splitting of the less plastic layers (or dykes and sheet intrusions) into blocks and other evidence of the flattening of the strata.

A regular feature of the internal structure of metamorphic series is schistozity, that is a plane-oriented structure parallel to bedding. Strongly metamorphosed rocks do not show either flow cleavage cutting across the strata, or fracture cleavage. All the flat crystals in them are oriented parallel to the strata and their bends take part in the development of small folds and puckering which are in evidence everywhere. Often plane-parallel orientation of the crystals is accompanied by linear orientation, the long crystals usually being oriented parallel to the strike of the folds.

Intrastratal schistozity apparently is explained by the fact that in the process of deformation of a plastic series owing to their flowage the material is redistributed chiefly within the strata. The flowage

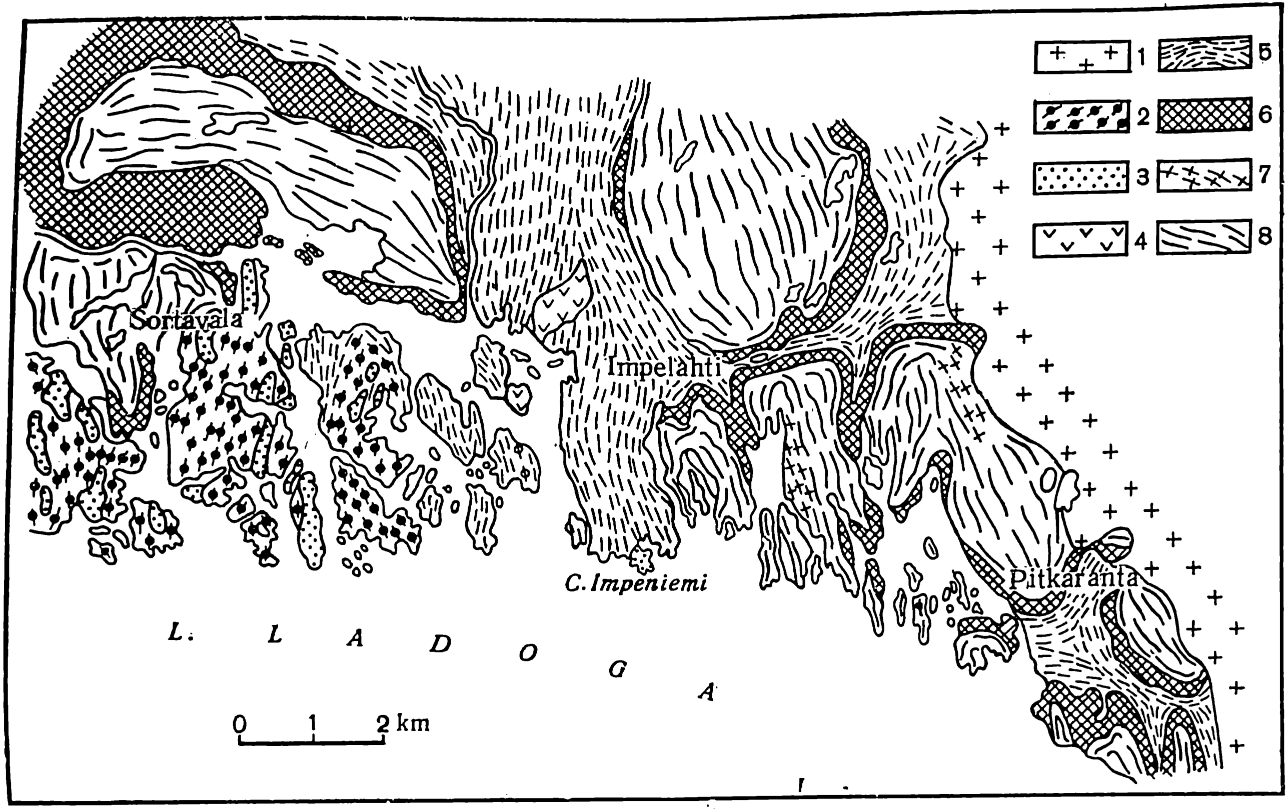


Fig. 106. Typical structure of Archean series. Gneiss domes fringed by metamorphic schists:

1—Jotnian rapakivi granites, post-Ladoga complex; 2—migmatites; 3—granites, granodiorites, diorites; 4—gabbro-diorites, Ladoga formation; 5—diorite-garnet, diorite-quartz schists and quartzites; 6—hornblende shales, amphibolites, Lower Archean complex; 7—gneisso-granites; 8—migmatized gneisses forming domes (after Eskola)

is parallel to the strata and takes place within them. Due to the different rate of movement of the material at different levels within a stratum and particularly between different strata flat crystals are turned parallel to the strata.

Some regional metamorphic sequences exhibit regional cataclasis, that is universal crushing of the grains of rocks which involves very thick series on vast areas.

Studies of oriented structures in igneous and metamorphic rocks belong to the domain of petrotectonics sometimes referred to as petrostructural analysis.

The object of petrographic study of rocks is to ascertain the dominant orientation of the minerals, the degree of their crushing and to explain the observed texture. The shape and position of crystals are investigated by means of the polarization microscope fitted with a Fyodorov stage for ascertaining the spatial position of the optical axes of crystals.

The dominant orientation of crystals is established statistically. It is known that tectonic deformation results in the development of a new texture. However, not all minerals assume new positions and by far not to an equal degree. Some attain it earlier, others later. Therefore, along with minerals that have already assumed a new position, there are some which have not attained it yet. In view of that measurements of the position of only a few crystals can yield only random results. This is why it is essential to measure the position of as many crystals as possible (hundreds and even thousands) and then, using statistical methods, obtain a true picture of the dominant orientations.

In regions of intense igneous activity and high-grade metamorphism very widespread are mixed rocks, migmatites, briefly mentioned earlier. In some instances these rocks have great structural significance. Migmatites are composed of two principal components: the host rock which is preserved either as thin layers or as separate fragments of various size and in different state of preservation, and the igneous material that fills the interstices between the fragments and consists of quartz, feldspars, granitoids, pegmatites or granite-gneisses.

Several varieties of migmatites are recognized, depending on the size and position of the remnants of the initial rocks, and the quantitative ratio between these rocks and the igneous material.

Foliated migmatites are formed as a result of intrusion (or are formed in the enclosing rocks as a result of their partial fusion) of pegmatitic, aplitic or granitic magma as thin intercalations between beds of intensely metamorphosed rocks of sedimentary origin but usually altered to gneisses or crystalline schists. In this case the bedding of the metamorphic series remains undisturbed.

Branching migmatites are formed by the injection of igneous material into the enclosing rock along a more or less complex network of small and big passageways (fissures and cracks) crossing the layers in various directions, but in some places coinciding with the bedding planes (Fig. 107).

Agmatites represent the next stage in the absorption of the rock by igneous matter. In this case the initial metamorphic rock survives as completely detached blocks, large or small, suspended in the

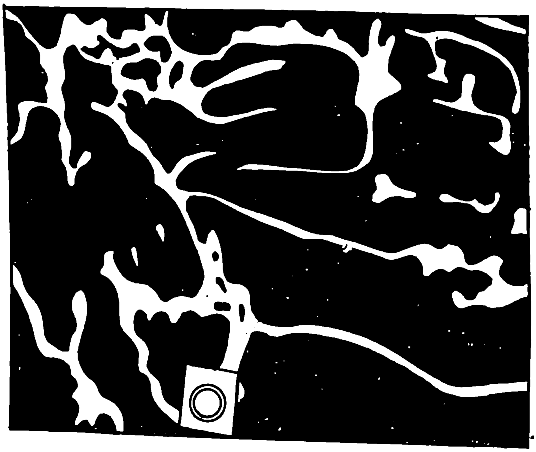


Fig. 107. Branching migmatite

igneous material. Most striking are agmatites whose initial dark-coloured rocks (amphibolites, for example) are included as huge fragments in a granite-gneiss, aplitic or granitoid groundmass (Fig. 108). In this case the fragments may be either in close contact or widely separated from each other. Sometimes the fragments show a certain regular arrangement. For example, it may correspond to the original lamination of the primary rocks. In this case usually there is some evidence showing that initially the separation of the fragments was associated with the splitting of certain layers into blocks and lenses. In other cases the parallel arrangement of the elongated fragments with a planar-oriented texture of the igneous rocks shows that the blocks have been turned around by the movement of the igneous material. The plane-parallel oriented textures of the igneous material often "flow around" the contours of the blocks.

Shadow migmatites develop at the last stage of the absorption of the original rock by the igneous material. At this stage the distinc-

tion between the original rock and the injected igneous material is almost completely obliterated. The blocks of the host rock are discernible but their outlines are very vague, being "dissolved", as it were, in the granitic magma. Often these blocks can be recognized only on the weathered surface of the migmatite (owing to non-uniform weathering), but are undistinguishable on fresh fracture. With further intensification of magmatic process, the migmatites pass into granite-gneiss or granite.

An interesting formation occasionally observed in metamorphic

Fig. 108. Agmatite

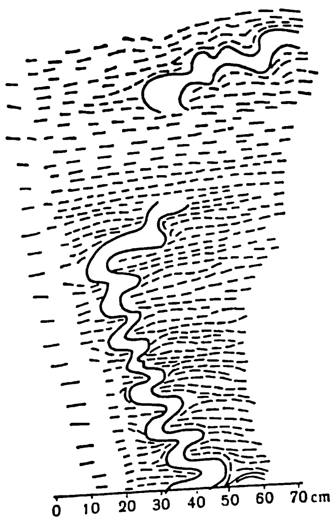


Fig. 109. Ptygmatite vein

series and migmatites are the so-called *ptygmatites* or *ptygmatite veins*—sinuous veinlets of granite-aplite, pegmatite or quartz. They are usually a few centimetres thick, with numerous convolutions tens of centimetres long that cut across laminations, layers and schistozity in all directions (Fig. 109). The origin of the convolutions is still uncertain. When such veins cut across schistozity it may be assumed that originally they were flat and constituted the filling of fissures. The convolutions were produced later as a result of sliding of the rock slabs along schistozity planes.

Another important problem is that of the effect of tectonic stresses on the physico-chemical migration of substances in the earth's crust. A very widespread phenomenon, for example, is the filling of tension fractures in rocks with calcite or quartz. Inasmuch as such fractures are often very small and bounded on all sides, it is perfectly obvious that the filling matter could derive only from the enclosing rocks. Therefore, the sharp pressure drop caused by the development of the opening results in selective absorption or "sucking in" of certain substances from the enclosing rocks. On a small scale this is the

process by which the matter of the earth's crust separates into substances of different composition.

Furthermore, it has been noted that maximum concentration of aplitic and quartz material occurs at those places where displacement (in the course of slipping of one part of a series on another) was greatest.

Without question, the phenomena of tension and compression in the earth's crust have a marked effect on the nature of the processes occurring in magmatic hearths. In the first place they affect the processes of differentiation of the magmatic material, of the assimilation of the enclosing rocks by it, and also the mechanism of intrusion.

However, the entire complex of the last-mentioned problems does not fall within the purview of structural geology as such and we mentioned them only to attract the attention of those geologists who study the structure of igneous and metamorphic rocks.

CHAPTER VIII

Structural Complexes

Structural forms of different kinds never occur singly. Always they appear in different combinations and the relationship between them is controlled by certain dependences. To reveal these dependences is one of the main tasks of geotectonics.

Combinations of structural forms develop under different conditions and are determined by various factors.

In some cases, for example, these combinations may be the result of the historical succession of events in the development of structural forms. This is to say that a certain segment of the earth's crust may have repeatedly suffered dislocations and in this way the different modes of occurrence of rocks may have been superposed on each other. In this case we may speak of a *historical combination of structural forms.*

Alternately, different structural forms may have developed at the same stage of structural development in generally similar physical conditions due to non-uniformities of the medium and deformation, and also as a result of natural passing of one kind of a dislocation into another (for example, the passing of plastic deformation into viscous fracture). In this way big and small folds and faults, etc., might have developed either simultaneously or in rapid succession in the process of a continuous deformation. This may be referred to as *mechanical combination of structural forms.*

Lastly, a combination of structural forms can be expressed on the regional plan. The distribution of structural forms and their complexes over the earth's surface is controlled by certain dependences, which in turn are determined by the general dependences that control the development of the earth's crust. This is known as *regional combination of structural forms.* We shall now consider some of these combinations.

28. HISTORICAL COMBINATION OF STRUCTURES

Historical combination of structural forms is expressed best in *unconformable bedding and stratigraphic intervals.* When the bedding of sedimentary strata is parallel and nothing indicates any long

interval in sedimentation such a mode of occurrence is referred to as conformable. It bears evidence to continuous subsidence of the earth's crust in the region of accumulation of the sedimentary rocks.

If, however, the accumulation of sedimentary strata was interrupted by uplifting of the earth's crust followed by renewed subsidence and accumulation, an interval in sedimentation and a stratigraphic hiatus are observed in the strata deposited before and after the uplifting. The duration of such an interval may vary widely. In some cases as a result of such an interval whole geological systems or stages may be missing from the section, in others they are so short that their duration cannot be established by paleontological methods,

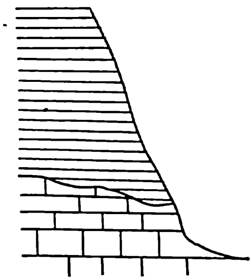


Fig. 110. Parallel unconformity

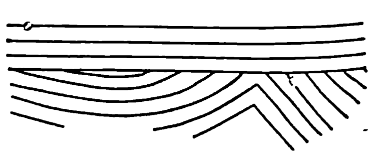


Fig. 111. Angular unconformity

since the outward appearance of the organisms did not change noticeably within this period. Such an interval is established only on the evidence of erosion within the sedimentary sequence, such as surface irregularities in the eroded part of the rocks, conglomerates with pebbles in underlying rocks, etc. Such short intervals recorded only in the structure of the sedimentary sequence, but undetectable by the paleontological method are usually referred to as *intraformational hiatuses.* They indicate that subsidence of the earth's crust during the accumulation of the given series was interrupted from time to time by short periods of uplifting.

If during the uplifting and the interval, the position of the strata did not change and they remained horizontal, the layers that are deposited after the interval will be parallel to those deposited before the interval, and will be horizontal too. Such a relationship between strata is referred to as *parallel unconformity* (Fig. 110).

In contrast to that, *angular unconformity* develops when during the interval the position of the earlier-formed strata changed, that is, they were tilted or crumpled into folds (Fig. 111). It is evident that the development of angular unconformity comprises the following steps: (a) accumulation of a lower horizontal complex;

(b) folding of the strata; (c) uplifting of the earth's crust; (d) erosion, as a result of which the newly-formed folds are partly destroyed, and folded strata become near-horizontal at the surface; (e) subsidence of the earth's crust, in the course of which an upper complex of horizontal strata is accumulated.

Angular unconformity may be characterized by the angle of unconformity, that is the angle between the dip of the strata in the lower series and the dip of the surface of unconformity, that is of the foot of the overlying complex. From one place to another this angle may vary within wide limits.

In Fig. 111 the angle of unconformity increases over the curve of the anticline of the lower complex and diminishes over synclines. Over the axis of the syncline it may fade out altogether. This apparent conformity in the bedding of the strata belongs to the type referred to earlier as parallel unconformity. However, if the anticlines of the lower complex have broad arches, such a decrease of the angle of unconformity may be observed over the axes of anticlines as well. Then the angle of unconformity remains most pronounced over the limbs of the folds of the lower complex.

If angular unconformity is connected with a change not only in the dip of the strata but also with a change in their strike, such unconformity is referred to as *azimuthal*.

In some instances, particularly on platforms, the angle of unconformity is so small that it cannot be detected directly in exposures. It can be revealed only by tracing the contacts of different series over large areas. Then it becomes evident that the foot of a certain series gradually, over many tens and even hundreds of kilometres, cuts across the underlying series, coming into contact now with older, now with younger series. Such unconformities characterized by very low angles are referred to as *geographic unconformities*.

The surface of unconformity most often is smooth, though occasionally it may be dissected and retain some features of the ancient relief formed during the era of denudation. Such irregularities are usually negligible and do not exceed a few metres, or rarely a few tens of metres in height.

Deposition of sediments on an irregular surface has a number of peculiarities. When the irregularities are gentle this may result in *blanket-like deposition* in which the thickness of the lower layers of the overlying series varies with the irregularities of the surface of unconformity, diminishing over projections and increasing over depressions. Deposition over a more rugged relief, results in a so-called *overlap* (Fig. 112) which may be *parallel* or *unconformable*.

The angular unconformity shown in Fig. 111 has formed in one stage. The upper sequence deposited after a break was not dislocated and retained its horizontal bedding. However, angular uncon-

formities may reoccur many times in the vertical section corresponding to repeated uplifts and dislocations. It is evident from the section shown in Fig. 113 that the strata of the second complex from the bottom deposited unconformably on the dislocated lower complex were themselves crumpled into gentle folds. A later upwarping

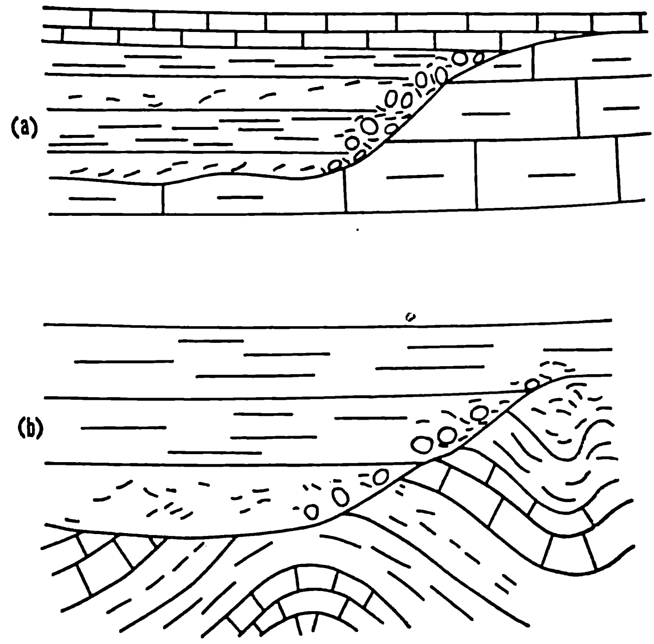


Fig. 112. Overlaps:
a—parallel; *b*—unconformable

resulted in the appearance of a new surface of unconformity, on which a third complex, which remained horizontal, was deposited.

The section also shows tectonic faults of two generations. One of them involved the lower complex, but had developed before the deposition of the second complex; the second had developed after the deposition of the two complexes but before the deposition of the third.

These repeating angular unconformities divide the section into structural stages, which differ from each other in the degree and character of dislocations.

Figure 113 shows that the lower complex is folded more intensely than the upper second complex. This in part may be due to the fact that after the deposition of the lower complex folding was more

intense than after the formation of the second complex. Most important, however, is the fact that the lower complex suffered folding twice: after the deposition of its own strata and after the accumulation of the second complex, when it was folded again together with the latter.

Often the folded forms below and above the surface of unconformity may belong to different types. For instance, the folds of the lower

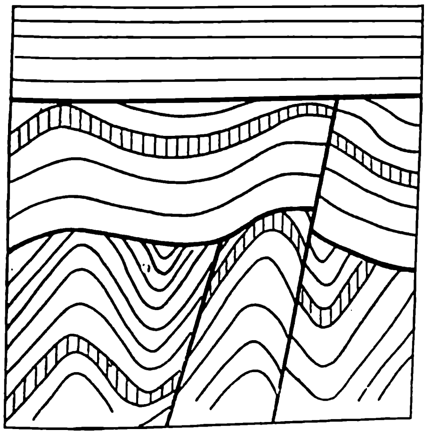


Fig. 113. Double angular unconformity. Tectonic faults of two generations

complex may be complete while the upper complex may comprise only discontinuous folds, or be characterized by gentle syneclises or anteclises alone (Fig. 114). Hence, the description of any mode of occurrence of rocks never fits the entire section of the earth's crust, but only a more or less limited sequence beneath and above which different structural forms may be found.

It is evident from the above examples that the most "pure" structural forms which have retained their original appearance occur in those sequences that were dislocated only once. On repeated dislocation, in some (deeper) series later dislocations are superposed on earlier ones.

This may greatly change the outward appearance of structural forms. For example, Fig. 115 shows two series separated by an angular unconformity. When the deposition of the upper complex began, its foot, which at the same time constitutes the surface of unconformity, was horizontal. Projecting it mentally into the horizontal position, we shall see that by the time the accumulation of the upper complex began, the anticlinal bend in the lower complex did not

exist: the strata of the lower complex in the right-hand part of the section were tilted to the right, while in the left-hand part they were horizontal. Thus, at that time the lower complex was not an anticline but a flexure. It was only later, after the deposition of the second upper complex, that the strata of the lower complex were dislocated

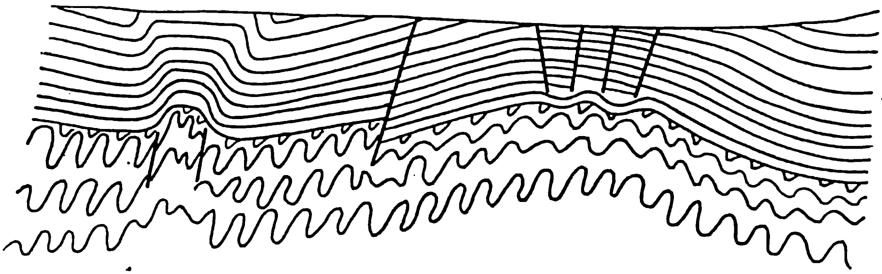


Fig. 114. Two structural stages with different types of dislocation

once more, upon which the left-hand part of the complex was tilted to the left. Thus, it is evident that the anticline in the lower complex was formed not in one, but in two stages and that the right limb was formed before the left.

This is an illustration of the method commonly used to reconstruct the tectonic history of a region in which angular unconformities

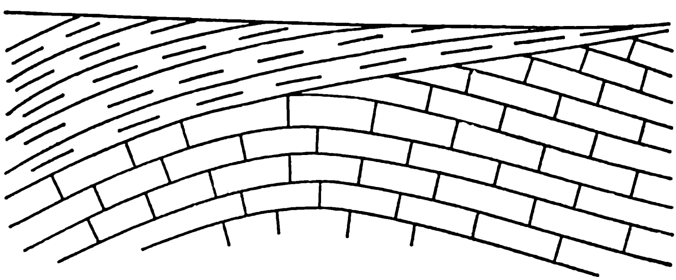


Fig. 115. Section attesting two-stage formation of an anticlinal fold in the lower complex

are observed. The substance of the method is that the surfaces of angular unconformities are mentally or graphically projected onto the horizontal position. Simultaneously we reconstruct the structure that existed by the time when the deposition of the next complex overlying the unconformity began.

If, for example, we project onto the horizontal position the lowest surface of unconformity shown in Fig. 113, this would give us the

structure that existed when the deposition of the second complex of sediments began from the bottom upwards. The "straightening" of that surface of unconformity would require a certain flattening of the folds in the lower complex since the angles of unconformity between the strata of the lower and the second complexes must be preserved. Thus, it would be established that when the deposition of the second complex began the folds of the lower complex were more gentle. They became steeper when the lower complex was dislocated once more together with the second complex.

It will be seen from Fig. 114 that the lowering and uplifting of the general level of folds in the lower complex went along with dislocations of the upper complex and did not exist before the latter was deposited.

In such operations one should also take into consideration the existence and age of tectonic faults. Reconstructing, for example, the structure as it was when the deposition of the second complex began (Fig. 113), we should bear in mind that at that time right-hand fault did not exist yet; since it developed after the accumulation of the second complex across which it cuts with the same throw as in the lower complex. However, had the throw along this fault been greater in the lower complex than in the upper, the natural inference would be that the fault developed in two stages: first, in the lower complex when its throw was smaller than it is now, and later after the deposition of the second complex, it was "revived", extended to the second complex, and the displacement along it was repeated. The net throw of this displacement can be seen in the second complex. In the lower complex it was added to the throw of the first movement along the same fault.

29. MECHANICAL COMBINATIONS

Examples of mechanical combination of structural forms are many. For instance, folding of strata is often combined with boudinage. We know that folding and boudinage are mechanically interconnected and constitute a common combination of structural forms.

Figure 89 shows a fault some beds of which pass into a flexure. From what has been said above concerning faults and flexures it is evident that these two dislocations are mechanically interconnected—a flexure as it develops gradually passes into a fault.

More complex combinations of folds of different orders and faults of different magnitude are shown in Fig. 116.

All these dislocations developed together in the course of a single deformation process and were determined by the non-homogeneity of the medium. They are recorded in the superposition of folds and faults of different orders.

The general dependence governing the mechanical combination of structural forms is that the forms involved in a given combination occur as different but associated manifestations of deformations conforming to a single plan which continue or supplement each other. This principle of "association" is extremely important for establishing whether a given case is a mechanical combination of structural

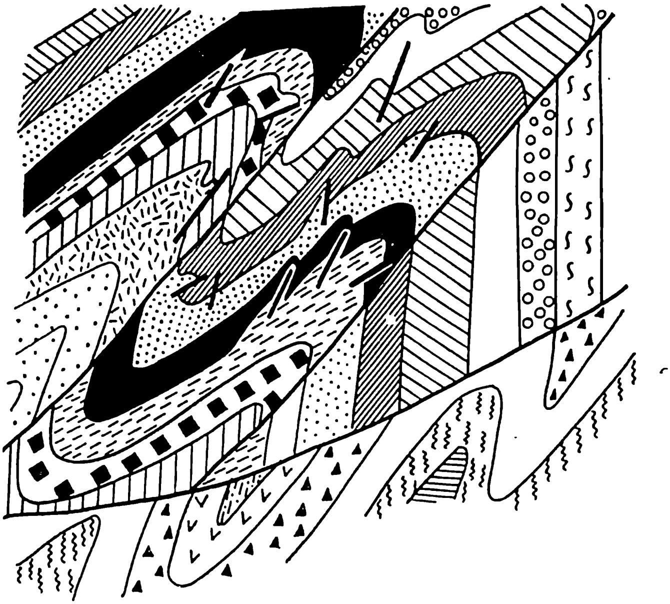


Fig. 116. Example of mechanical combination of structural forms: folds and associated thrust faults of different orders

forms produced in the process of a single deformation, or it is historical, in which structural forms of different types developed in different conditions that changed with time.

A few more examples are in order. The development of upwarps and downwarps in connection with vertical differential uplifts and subsidences of the earth's crust is accompanied at those places, where the movements were sharply pronounced, by flexures which may pass into normal or reverse faults. Naturally, such faulting occurs when the amplitude of the downwarping or upwarping is great enough. This is why arch uplifts and depressions are so often combined with normal and reverse faults. In such cases the movements along the faults supplement and continue the upward and downward bending of the earth's crust due to which these dislocations may be regarded as associate. Similar combination of plastic bending with

faulting is observed in connection with the development of intermittent, particularly box-like folds, whose marginal flexures often pass into faults.

In the case of local upwarps and other uplifts of the earth's crust (anteclises, arch uplifts, discontinuous folds) other combinations of plastic bending and faulting may be observed as well. It is known that any local upwarp of the earth's crust generates tensile stresses, which may cause the development of normal faults and sagging of the crest of the uplift. From the geometrical point of view these are

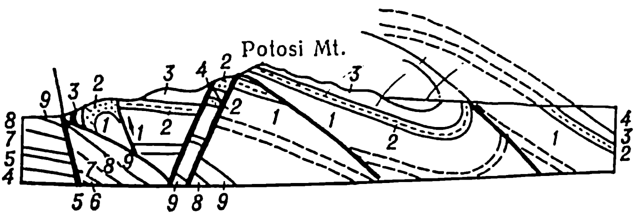


Fig. 117. Profile showing a case of subsequent dislocation of folds and overthrusts by faults (thick lines):
1—Upper Cambrian-Devonian; 2—Devonian; 3-6—Carboniferous; 7-8—Triassic; 9—Jurassic

reverse faults since they result in downthrusting of a segment of an uplifted region. From the mechanical point of view, however, they are associate since they correspond to the conditions of tension that are dominant in the bent strata. A combination of complete folding with imbricate overthrusts was shown in Fig. 94. In this case, too, complete mechanical harmony is observed: the overthrusts continue and develop the deformation that originally was expressed in folds, i.e., the deformation caused by horizontal compression.

Examination of the structure shown in Fig. 117 leads to somewhat different conclusions. Here a folded structure, though accompanied by imbricate overthrusts, is furthermore dislocated by a series of normal faults. It is evident that no mechanical association exists between faults and the folds with overthrusts since the latter reflect conditions of compression, and the former conditions of extension. From this it may be inferred that the tectonic conditions in the given region had changed and were first characterized by compression which resulted in the development of folds and overthrusts and later yielded place to extension, which resulted in normal faults.

Proper understanding of mechanical combination of structural forms helps to avoid errors in regional geologic surveys. For example, if one observes in regions of development of discontinuous folds outcrops of tectonic faults, whose angle of dip cannot be established, the natural inference would be that these are normal or reverse

faults but not overthrusts, since these do not combine mechanically with discontinuous folds which are produced by the action of vertical tectonic forces. Similarly, the crest of a box fold of block origin can hardly be expected to contain additional recumbent folds with a reversed position of the strata.

The above examples enable a more precise definition of the conception of mechanical combination of structural forms. To be more specific, it is not always that dislocations which are found in a mechanical combination have developed simultaneously. We may even say that practically always they do not develop simultaneously but succeed one another. An important point, however, is that they are connected with the development of a single deformation, reflecting common mechanical conditions in the earth's crust, such as compression, upwarping, etc. What is meant by historical combinations, as different from mechanical ones, are such structural forms that have developed successively and also under different mechanical conditions or with a substantial interval in the development of deformations (usually recorded in angular unconformity).

30. REGIONAL COMBINATIONS

A regional combination of structural forms constitutes, as it were, the supreme expression of structural relationships. It reflects the most general and fundamental dependences governing the tectonic movements of the earth's crust. This problem in its entirety is the domain of geotectonics and we mention it here only in passing.

It is known that there exists a dependence, expressed in the division of the earth's crust into regions characterized by tectonic movements of different intensity and nature. The two principal types of such regions are platforms and geosynclines. On platforms tectonic movements are quiet, their intensity is low, while in geosynclines the tectonic life of the earth's crust is very intense. Similarly, mag-

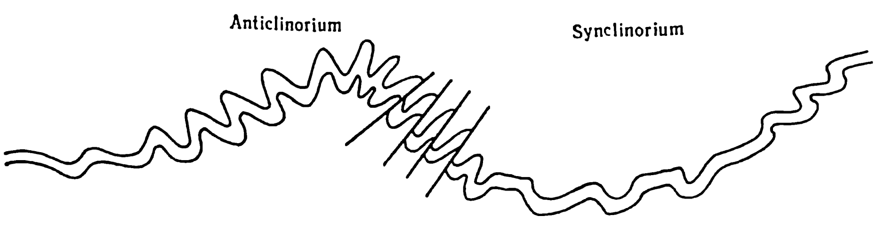


Fig. 118. Idealized schematic of the occurrence of rocks in an anticlinorium and synclinorium

matic processes are much more active and diversified in geosynclines than on platforms.

In view of the above the structures of platforms and geosynclines substantially differ between themselves.

On platforms the strata are nearly horizontal over large areas and their bedding is disturbed only by gentle and extensive arches of anteclises and equally gentle and vast troughs of synclinal downwarps. Discontinuous folds, chiefly of block origin, which are extensively developed on platforms, have a greater disturbing effect. Most often they constitute reflected block folds, expressed as very gentle domes and brachy-folds. Occasionally, however, they are more sharply pronounced as box folds of great throw with steep limbs. These most often occur singly amidst platforms.

In some places injection folds occur on platforms as well. For the most part these are diapir salt domes which owe their origin to gravitational buoyancy.

Disjunctive dislocations occur on platforms mostly in the form of normal or reverse faults, associated with the uplifting of anteclises or with discontinuous folds. They are confined to these uplifts and reflect either their elevation or extension at their arches.

Continuous folding is not typical of platforms. Neither do great intrusions (batholiths) occur here. Only in rare instances small intrusions (fissure intrusions, stocks) and also sheet intrusions, occur on platforms. Rather profuse effusions are observed on some platforms. They mostly belong to the fissure type and are characterized by uniform basaltic composition (plateau basalts or traps).

Entirely different is the structure of geosynclines. Owing to intense and sharp oscillatory crustal movements a geosyncline is always divided into high uplifts and deep troughs whose relative throw sometimes amounts to 12 to 15 kilometres. The uplifts and troughs,

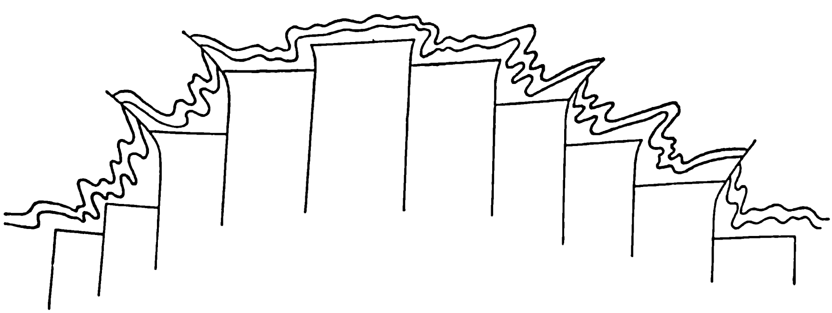


Fig. 119. Idealized schematic of the structure of an anticlinorium divided into step-like blocks

however, do not retain their "pure form" but are always complicated by intense folding, mostly of the continuous type on uplifts, while in depressions it is mostly intermediate (ridge or box type) and partly discontinuous, but much more intense than on platforms.

Uplifts complicated by overlying folded strata are called *anticlinoria*. Troughs complicated by folding are called *synclinoria*. Thus, structurally a geosyncline is divided into anticlinoria and synclinoria.

Like all uplifts and troughs, anticlinoria and synclinoria are elongated and oval and extend for hundreds of kilometres. In either direction along the strike they terminate periclinally and centriclinally. An example of an anticlinorium is the Main Caucasian Range; an example of a synclinorium is the Kura-Rion depression that adjoins it in the south. On anticlinoria continuous folds as a rule are overturned fan-like away from the axis of the anticlinorium towards the adjacent synclinoria and often are combined with overthrusts, whose position and direction of movement correspond to the inclination of the folds (Fig. 118).

Because the strata in a geosyncline are everywhere crumpled into folds, it is also referred to as a folded region or a folded zone. At present, the term *folded zone* is used most commonly to describe the structure of a geosyncline, while the term geosyncline proper is used to characterize the nature of the development of the earth's crust, i.e., the complex of intense tectonic and magmatic processes (geosynclinal type of development). Inasmuch as here we shall speak mainly of structural features, we shall dispense with the term "geosyncline" in favour of "folded zone".

Apart from folding and imbricate overthrusts, anticlinoria may be disturbed mainly by longitudinal, but sometimes also by transverse and diagonal associate faults of normal or reverse types as a result of which the structure of the anticlinorium becomes step-like. Such steps formed by longitudinal faults are best expressed in the cross-section of the anticlinorium (Fig. 119). Transverse faults which usually develop at the periclines generally impart a step-like appearance to the plunges of the anticlinorium in its longitudinal section.

Anticlinoria may also be disturbed by faults of different systems, which are not combined mechanically with crustal upwarps, but divide the anticlinoria into blocks displaced vertically relatively to one another.

Folding at the limbs of an anticlinorium is usually non-uniform with local intensifications and attenuations. Often most intense and complicated folding is observed beneath the ledges of uplifted blocks that divide the limb of the anticline into steps. Such zones of intense folding are referred to as *crush zones* (Fig. 120).

Gravity overthrusts or nappes descending into adjacent synclinoria are often associated with the limbs of anticlinoria.

The structure of synclinoria is always more quiet than that of anticlinoria. As noted earlier, intermediate and discontinuous folding is most common in synclinoria.

Folded zones are characterized by numerous big and small magmatic manifestations. These are concentrated mostly in anticlinoria. Batholiths—enormous granite masses confined, as a rule, to the axial parts of anticlinoria are found only in folded zones. Fissure and sheet intrusions are also extensively developed in such zones. The first cut across the batholith and the enclosing folded stratified

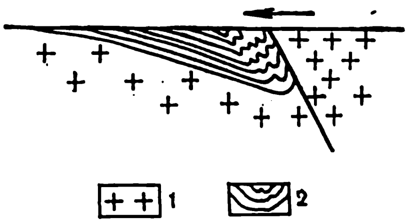


Fig. 120. Schematic of the zone of crumpling in Paleozoic deposits of Central Kazakhstan (after Y. Zaitsev): 1—basement; 2—sedimentary Paleozoic rocks

rocks in various directions, while the latter lie between the strata and are crumpled into folds together with them. Manifestations of effusive magmatism are observed in such regions as well. Some ancient lava sheets, similarly to sheet intrusions, occur within the sedimentary strata and are crumpled into folds, whereas other lava flows unconformably overlie the folded series.

Magmatic manifestations are closely associated with extensive metamorphism which involves numerous series of the anticlinorium.

The character of magmatic manifestations in the lower parts of the limbs of anticlinoria and synclinoria is somewhat different. The intensity of magmatism is much lower, and it is expressed mostly in small stocks, laccoliths and fissure intrusions. Synclinoria, however, are often completely devoid of magmatic manifestations.

It is evident from a comparison of the structure of different regions that an intermediate complex should be recognized between regional structural complexes of the platform and the geosynclinal types. This complex is less complicated than a typically geosynclinal complex but more complicated than a typical platform complex. Regions characterized by intermediate semi-geosynclinal, semi-platform structural conditions are referred to as *parageosynclines*. Within their limits folding is of an intermediate nature and magmatic activity is generally expressed in small intrusions and few effusions.

It has been mentioned earlier that synclinoria in folded zones are characterized mostly by intermediate structural conditions. Such

conditions are also observed at the periclines of anticlinoriums (in the regions where they plunge and terminate). Examples are the northwestern and southeastern plunges of the Main Caucasian Range, where intermediate, chiefly ridge-like folding replaces the continuous folding of the central parts of the range. In turn, at the edges it passes into discontinuous (mostly diapiric) folding of the Taman Peninsula in the northwest, and of the Apsheron Peninsula in the southeast.

Typical intermediate zones are the so-called *fore deeps*, which practically always are located between folded zones and platforms. They are deep (the sedimentary series are 4 to 5 km thick) tectonic depressions. Their structure is asymmetric; the inner limb adjoining the nearest anticlinorium is more steep while the platform limb is wider and more gentle. The limbs of a fore deep are often dislocated by step faults and flexures. On the folded zone side, continuous folding sometimes partly penetrates into the fore deep and farther. Within the fore deep it is replaced by intermediate folding and this, in turn, by discontinuous folding. Intermediate folding is most characteristic of fore deeps, but diapiric salt domes are also widely developed here. Magmatic manifestations are very rare, and are represented by small and fissure intrusions.

Earlier we characterized anticlinoria as arch uplifts complicated by folding. From this point of view it is interesting to note the existence of crustal upwarps, which in part are simple arch uplifts and in part anticlinoria. An example is the Greater Caucasus. Over a long stretch in the middle part of the range (from the Belaya R. in the west to the Chegem R. in the east) the tectonic structure of the northern slope of the ridge is very simple. This slope constitutes a gentle (10° at most) smooth monocline composed of Mesozoic and Cenozoic strata, entirely free from folding and complicated only in some places by small local flexures, which means that this is a typical limb of a simple arch uplift. However, the axial part of Greater Caucasus, particularly its southern limb, is complicated by most intense complete folding and, therefore, this zone of Greater Caucasus is a typical anticlinorium (Fig. 121). The southern limb of Greater Caucasus is steeper than the northern limb and more intensely folded.

Every regional combination of structural forms (geosynclinal, platform and parageosynclinal) is always confined to a definite stratigraphic complex, to a definite structural stage. Therefore, the concept of regional structural complexes is meaningful only as applied to a certain structural stage occupying a definite stratigraphic position.

Thus, when we spoke earlier of anticlinoria and synclinoria, we had in view the occurrence of Jurassic, Cretaceous and Tertiary

rocks. These rocks compose a single structural stage and jointly correspond to a definite major phase in the development of the earth's crust in this region. Older rocks are separated from this complex by regional angular unconformity and are characterized by a different structure, in which the anticlinoria and synclinoria are differently distributed.

In the case of the Russian Platform, a single structural stage is formed by rocks from those of the Riphean system to Tertiary rocks higher up, which comprise an enormous time interval from the Early Paleozoic to recent. It is in this rock complex that are observed those anteclises and syneclises and also the most diverse discontinuous folds, which together form the platform structure. Lower down lies a Precambrian complex of an entirely different nature. It is composed of crystalline and strongly metamorphosed (intensely dislocated) rocks which correspond to an entirely different stage in the development of the earth's crust. That stage preceded the platform stage and generally may be described as geosynclinal. Such two-stage structure is characteristic of all platforms without exception. The upper platform stage, with structural forms typical of platforms, overlies with sharp unconformity a lower, strongly dislocated and metamorphosed, older stage, usually referred to as the *folded basement* of the platform.

All the problems of regional combination of structural forms discussed in this chapter are treated more fully in corresponding courses of geotectonics. This establishes a number of historical dependences governing the development of tectonic movements, which is essential for understanding the regular features observed in the regional distribution of structural forms.

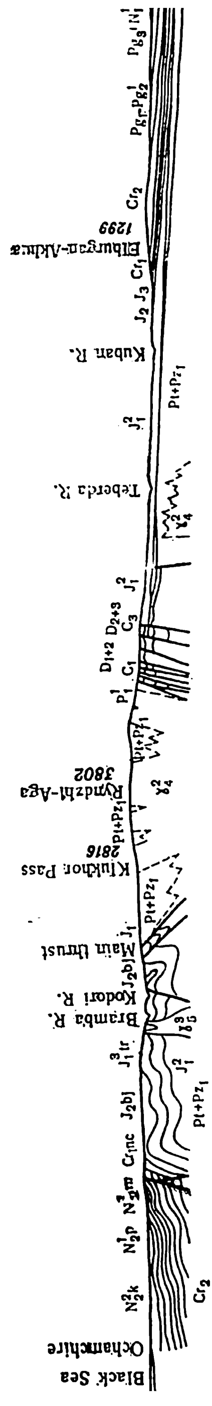


Fig. 121. Profile of the Main Caucasian Range (across the Klukhor pass). Monocline in the north; block structure in the axial zone; folds on the southern slope

CHAPTER IX

Non-Tectonic Dislocations

Various exogenous processes often result in local disturbances of the attitude of rocks occurring at shallow depth. The effects of these phenomena and processes must be studied primarily to avoid mistaking non-tectonic dislocations for tectonic ones and also to avoid errors in the interpretation of the tectonic structure of a given area. Geological records often contain serious errors of this kind due to the fact that certain surficial causes of dislocation of rocks have been ignored. Such mistakes are particularly common in regions with scanty exposures. We shall now briefly examine the most important non-tectonic factors and processes that disturb the attitude of rocks.

31. NON-UNIFORM COMPACTION

Slight dislocations of sedimentary rocks expressed in the form of very gentle dome-like uplifts and brachy-folds, and equally gentle troughs between them may be caused by non-uniform compaction of sediments under their own weight. It is known that in the process of their transformation into rocks certain sediments suffer considerable compaction, in the course of which they lose a part of the water they contained originally. This refers chiefly to clays. Freshly deposited silt contains 70 to 90% water by volume; compacted clay retains only 30% of the water. Sands and limestones are compacted to a lesser degree.

If clay was deposited on an uneven relief, composed of hard uncompressible rocks and its surface was levelled by the movement of the water, as a result of subsequent compaction the surface of the clay series will subside more strongly over the depressions of the basement than over elevations. This will produce "folds" that rapidly flatten out upwards (Fig. 122). Even in the absence of irregularities in the surface of the basement, such *supratenous folds* may develop due to the presence of isolated lenses of harder rocks, such as sandstones, limestones, etc., in the clay series. In this case compaction will result in greater subsidence of the clay layers in the intervals between the lenses, while over the lenses the layers will be convex.

This is why we may lawfully assume that certain extremely gentle archings and saggings of strata on platforms whose section is composed mainly of clays may have been produced by this mechanism. Supratenous folds are interesting to the oil geologist since they favour the accumulation of oil and gas. On the whole, however, folds of this origin are unimportant in the structure of platforms, since undoubtedly the folded structure of platforms is controlled chiefly by tectonic dislocations of the idiomorphic fold type.

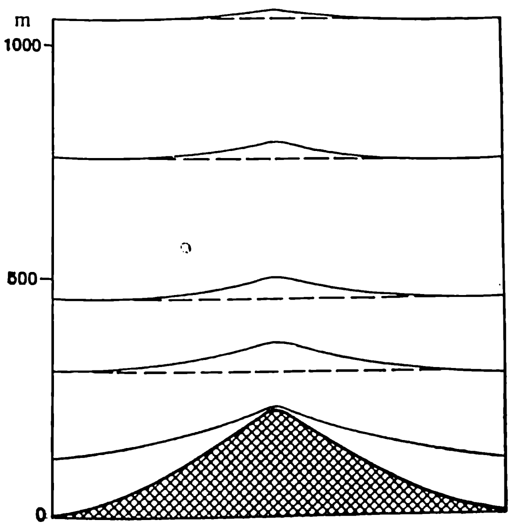


Fig. 122. Non-uniform compaction of clays over a projection of a rigid basement:
Basement—hachured; thin lines—positions of surfaces in the clay series (formerly horizontal) after compaction

Non-uniform compaction of soft sedimentary rocks (clays), with corresponding bending of the strata, may be caused by the non-uniform weight of the overlying rocks. Clays, for example, often form a *compaction basin* under the local weight of an overlying isolated reef mass of coral limestones.

32. CHANGES IN THE VOLUME OF ROCKS DUE TO PHYSICO-CHEMICAL PROCESSES

It is known that under the action of subterranean waters anhydrite passes into gypsum increasing in volume by 60%. Once this was believed to be the mechanism responsible for the formation of

diapir domes with gypsum cores. It was also assumed that other rocks (salts, in particular) occurring near the surface, upon a change in their volume due to recrystallization, may cause considerable dislocations in the attitude of the enclosing rocks.

It is now known, however, that in nature the transformation of anhydrite into gypsum occurs at a depth of not more than 150 m. At that depth, large anhydrite masses experience (usually as a result of their elevation in the process of formation of diapir cores) the effects of factors that were examined above. Thus, the increase in rock volume is only an additional, not the decisive factor, which causes the diapir core to rise to the surface.

If the anhydrite occurred between beds of other rocks, such as clays, as a result of the increase in volume occurring when anhydrite passes into gypsum, the anhydrite layers are usually crumpled into small folds or fractured and converted into breccias. Fragmentation and brecciation of intercalations of hard rocks between bands of anhydrite that undergoes hydration is observed as well.

33. ENVELOPING

It was said in the first chapter of this book that as a rule the original bedding of sedimentary strata is horizontal. Some exceptions to this rule were also noted. Furthermore we returned to this question when we dealt with unconformities. In some cases, due to irregularities in the sedimentation surface, the sediments form layers that are bent upward or downward. Such primary bending is always small in scale and confined to thin series; upwards along the section the bends rapidly level out, but nevertheless should be reckoned with.

Fig. 1 shows such primary inclined bedding of strata on the slopes of a Lower Carbon reef limestone mass.

Furthermore, enveloping has been recognized on some very gentle idiomorphic folds (domes) in the eastern regions of the Russian Platform. Figure 123 shows a series of nearly-horizontal Devonian and Franconian strata that were eroded and scoured. As a result there was formed an asymmetric hill, parallel to whose surface Famennian and Lower Carbon strata were deposited. Therefore, the dome-like occurrence of Upper Devonian and Lower Carbon strata is clearly due not to tectonic movements but to the enveloping of the buried relief. If should be noted, however, that in this figure the vertical scale is greatly exaggerated relatively to the horizontal. The boreholes *2* and *3* are separated by more than 4 km whereas the amplitude of buried relief does not exceed a few dozens of metres.

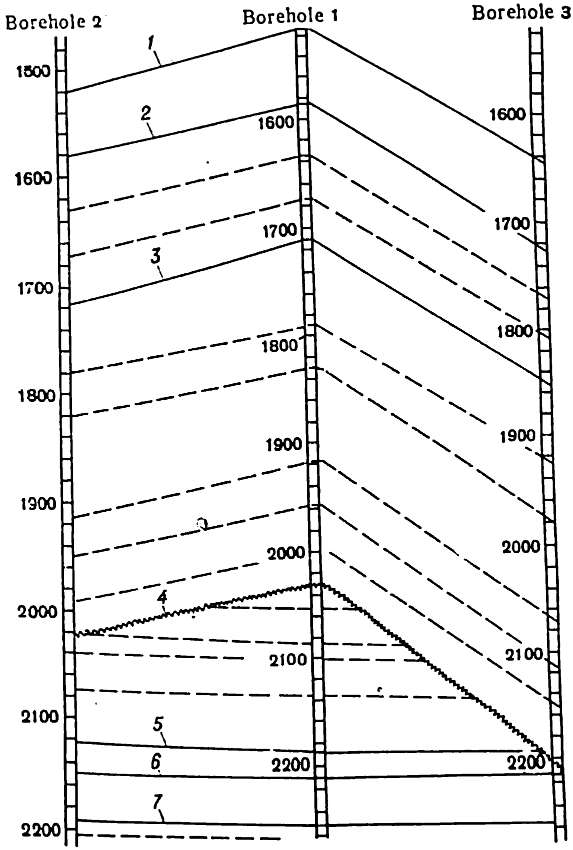


Fig. 123. Enveloping fold. Lobanovo area of the Volga-Urals region (after L. Rozanov).
Roof of: 1—coal-bearing horizon; 2—Tournai stage; 3—Devon; 4—surfaces of Franconian rocks; 5—domanik; 6—Paschian strata; 7—Givetian stage

34. SQUEEZING UNDER NON-UNIFORM LOAD

In our earlier discussion of the conditions of formation of injection folds we assigned the principal role to the movement of plastic rocks under the action of non-uniform weight of superincumbent strata.

Similar processes occur near the earth's surface but on so small a scale that it is difficult to regard them as tectonic phenomena. It will be easily understood, however, that the line between tectonic and non-tectonic processes of this kind is not easily drawn.

In Japan large quantities of volcanic [illegible] around one of the volcanoes. As a result of non-uniform [illegible]

of these volcanic products in those places where the load is less, the clays are squeezed upwards, forming small surface injection folds or clay diapirs.

Another frequent phenomenon is the heaving of plastic clay rocks at the floor of river valleys, as a result of their squeezing from beneath the high bank, whose rocks exert greater pressure on the clay series (this process in its more general form was considered in connection with injection folds). In this case too, however, the division between tectonic and non-tectonic processes is one of convenience and is determined by the scale of the phenomenon and its general effect on the tectonics of the given district. The movement towards the valley and heaving are confined to the plastic series and do not involve the underlying rocks.

To the same class belong the so-called neptunic dykes. These are intrusions of plastic rocks (clays, moist sand, or quicksands) into fissures in the overlying rocks. Upon induration these intrusions appear as bodies similar to dykes but are composed of sedimentary material. Generally such intrusions are produced under the load of the overlying rocks. Neptunic dykes resulting from the injection of water-logged Danian sands into fissures of Lower Syzran charts are known in the middle reaches of the Volga.

35. WEATHERING

In the process of physical weathering rocks that outcrop to the surface are subjected to cracking. As weathering proceeds the cracks become more and more numerous and those that had formed earlier grow wider. The first cracks to open in the course of weathering are primary fissures that had developed in the rock regardless of weathering as general tectonic joints. In a fresh rock these joints are often almost imperceptible, but are opened upon weathering. Upon continued weathering as more and more cracks are formed, the strike and dip of the new cracks is similar to those of the sets of joints that existed in the fresh rocks but were much fewer. Thus, in the fresh rock, in addition to the visible joints there exist parallel potential cracks or weakened directions which are revealed upon the development of joints produced by weathering. However, there are some exceptions to this rule and joints of different directions may develop upon weathering as well.

36. BENDING OF STRATA ON SLOPES

This variety of non-tectonic dislocations is of special interest, since failure to take it into account in many instances has led geologists to major errors. Bending of layers along slopes is observed

chiefly in mountain regions with steep valley sides and for the most part in relatively soft rocks, such as clays and clay shales. On a smaller scale, however, it is also observed in some small gulleys and in harder rocks (sandstones, and limestones).

The substance of this phenomenon is that in the zone of weathering the ends of strata cut by the slope bend over downslope under the action of gravity, which causes the slow downhill creep of hillside

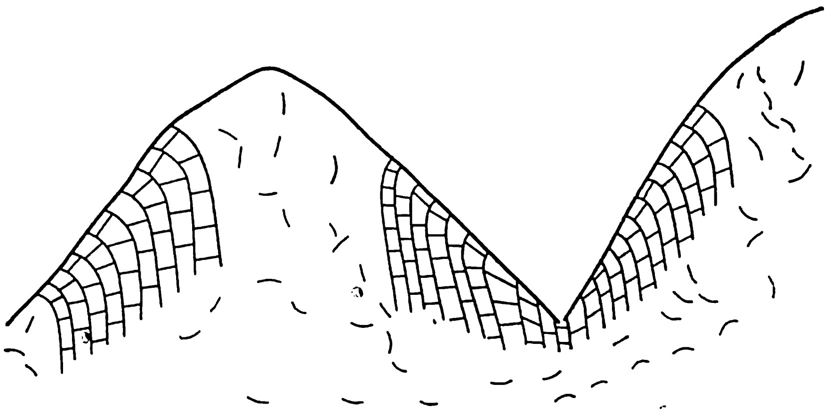


Fig. 124. Downhill bending of strata under the action of deluvial creep

waste. This often results in disturbance of the primary bedding and changes not only the dip angle but also the direction of the dip of the strata (Fig. 124).

In areas characterized by the development of clays and clay shales this phenomenon is widespread, in view of which the geologist is advised to display extreme caution in measuring the bedding elements. First he must convince himself that this bedding is primary and not false. The only reliable method in such areas is to measure the bedding elements of rocks along the thalweg, at the floor of gorges and in stream channels, that is in places as yet unaffected by the downward creep of hillside waste.

37. KARSTING

Karst results in the formation of cavities in limestone, dolomite, gypsum and salt series, which often causes subsidence or caving of the roof over the cavities which, in turn, disturbs the bedding of the roof rocks. With time the karst sinks and swallow holes that formed in this way are filled with fragments of the enclosing rocks, which gives rise to karst breccias. Such buried karst breccias are often found

in limestones of ancient geologic epochs. The breccias may contain large limestone fragments and even blocks comprising whole series of bands which may look like a primary occurrence.

Flexure-like downward bending of strata is observed occasionally around the edges of sinks and swallow holes.

38. LANDSLIDES AND ROCK FALLS

Landslides constitute a very active factor of non-tectonic dislocations both on the sea floor, in which case they are referred to as slumping, and on land. We had mentioned submarine slumping in the first chapter. Slumping sometimes occurs practically simultaneously with the deposition of sediments and involves material that has not been compacted or indurated. In other instances, however, judging by the fragments, the slumping involved indurated rocks (for example, hard limestones, which in the process of movement were broken into fragments).

Subaqueous slumping of silty layers is evidently possible even when the gradients do not exceed 1 to 2°. Very common is underwater slumping which involves series of layers a few metres thick on areas of hundreds and even thousands of square kilometres. There is evidence of underwater slumps that involved rocks hundreds of metres thick on areas of thousands of square kilometres.

In landslides, soft rocks are crumpled into small irregular folds and harder ones are crushed into landslide breccia. Expressed on a great scale, these phenomena were repeatedly mistaken for overthrusts and nappes with complex repetition of stratigraphic sequences. In some instances, even new phases of folding were established and angular unconformities were established where none existed.

Landslides are characteristically confined to a definite sequence of layers, above and beneath which these phenomena are not observed. A most characteristic point is the extreme irregularity of dislo cations associated with landslides.

Submarine slumps are observed chiefly in geosynclines. They are most often associated with certain stratigraphic horizons, which seems to indicate that certain periods were more favourable to submarine slumping than others. Such a favourable circumstance could be the increasing steepness of sea floor gradients due to greater intensity of oscillatory movements, which raised some sections of the floor relatively to others between regions of uplifting and erosion and regions of subsidence and accumulation. Earthquakes too play a very important role in subaqueous slumping, shaking up the sediments and thus causing their displacement. Both these factors which favoured subaqueous slumping were confined mostly to places and epochs of intense tectonic activity, expressed chiefly in sharper

up and down movements. From this point of view studies of subaqueous slumping are of definite interest to tectonics.

A variety of subaqueous slumping is the sliding of huge blocks and cliffs of hard rocks (mostly limestones) down gentle silty slopes under the action of gravity. As a result of such sliding of blocks over a slippery clayey sea floor, they may move for tens of kilometres from the place of their original occurrence.

Such, for example, is the origin of the so-called Dibrar cliffs in southeastern Caucasus—great (hundreds of metres) cliffs of Upper Jurassic limestones occurring in clays of the Barremian stage. These cliffs broke off the limestone bluffs that towered over the Barremian sea and then slid along the inclined floor for 20 to 30 km. Such movement could have occurred even in the absence of a continuous gradient of the sea floor all along the route from the precipice to the place where the cliff was buried. It could have been gradual, occurring in separate spurts, as an appropriate gradient developed at the given place. Earthquakes too may have been an important factor in inducing sliding.

Naturally, from the tectonic point of view it is interesting to determine the direction of the movement of the subaqueous slump and the time when it occurred.

Landslides are common in clay rocks on slopes along the river valleysides, at the shores of lakes and seas and in gulleys. The slump may involve areas of a few square kilometres, and cause the movement of rocks for hundreds of metres and even for a few kilometres downslope. The layer of rocks involved is usually not more than a few dozens of metres thick. The movement takes place along a curved surface which is steep in its upper part and gradually flattens out at the bottom.

Landslides are accompanied both by faulting and folding. The upper part of landslides experiences tension, and dislocations similar to tension faults develop there. In the lower part of the landslide where it meets resistance from irregularities of the relief, the rocks suffer compression which results in small folds and overthrusts.

Landslides produce breccias composed of angular fragments of the slumped rocks. Such is the probable origin of the breccia composed of fragments of Devonian limestones occurring in the Cretaceous deposits of the Altai. The breccia is unstratified and unsorted and more than 150 m thick in some places.

39. THE ACTION OF GLACIERS AND PERMAFROST

During their movement the Quaternary glaciers that covered a considerable part of Europe and America, in particular a great part of the European USSR, exerted pressure on the rocks over which

they moved, and in some places this resulted in folding and overthrusting.

Such glacial dislocations, represented by folds and overthrusts, are observed near Kiev along the Dnieper River.

Dislocations of this type are characterized by the fact that they extend only to the surface layers, to a depth of not more than a few dozens of metres and mostly involve Ancient Quaternary deposits.

Individual blocks moved by the glacier occur more frequently than folds. Great blocks of this kind may look like a primary outcrop and the attitude they assumed due to their displacement by the glacier may be mistaken for their primary attitude. For example, at the northern termination of the Polistovo-Lovat rampart where the bedding of rocks is entirely undisturbed, there is an outcrop of Silurian limestones that have been turned upside down and form precipices up to 26 m high. Boring revealed beneath the limestones a morain underlain by primary Devonian strata.

Specific non-tectonic dislocations develop in permafrost regions. In these regions the freezing of water which rises along fissures into the soil or issues as springs results in an increase in the volume of the soil. This causes heaving in the form of hummocks and hills, which sometimes contain laccolith-like masses of ice beneath the uparched soil layer. In view of their resemblance to laccolites they are sometimes referred to as *hydrolaccoliths*. The pressure of freezing water beneath the soil results in the explosion of such hydrolaccolith or permafrost hummock and the ejection of water, blocks of ice and mud.

Non-uniform thawing of permafrost sometimes causes subsidences resembling karst sinks. In other cases as a result of freezing of water in the cracks of bedrocks they are crumpled into fine folds by the pressure of the expanding ice.

SUBJECT INDEX

TO THE READER

Mir Publishers would be grateful for your comments on the content, translation and design of this book. We would also be pleased to receive any other suggestions you may wish to make.

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"Structural Geology" brings together in condensed form the fundamentals of the geology of primary and secondary forms of occurrence of rocks—sedimentary, igneous and metamorphic. It outlines the main principles of tectonic physics and sets out the basic mechanisms of folding and faulting. A classification of structural forms is given and the processes governing their historical, mechanical and regional combinations established. A special chapter deals with non-tectonic dislocations.
The book is intended as a textbook for students of geological faculties of universities, and schools of mining and geology.

VLADIMIR BELOUSOV, Corresponding Member of the USSR Academy of Sciences, the author of this book, is chairman of the International Committee for Project Upper Mantle. He leads a department of the USSR Academy's Institute of the Physics of the Earth, and holds the chair of dynamic geology at Moscow State University.

Prof. Belousov's main research has been directed to problems of geotectonics and of geophysics in general. He is the author of 200 published works, of which the best known are "Structural Geology", "Types and Origin of Folding", "The Earth's Crust and the Upper Mantle of Continents", and the monograph "Fundamental Problems of Geotectonics".

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